CHAPTER 3 — VISIBILITY

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3.1 Factors reducing visibility

Reduction in visibility depends on the concentration and size of hydrometeors, particulates or moisture in the atmosphere as well as the viewing path and the time of day. Factors reducing visibility are:
(i) Precipitation — rain, snow, hail.
(ii) Haze — condensation on dry particulates. Haze particles shrink and grow with changes in relative humidity (RH), high RH leading to poor visibility. Hygroscopic particles will encourage condensation at relative humidities well below 100% (3.9.2).
(iii) Haze in the dry atmosphere due to industrial pollutants and vehicle exhaust products — photochemical reactions can reduce visibility to <5 km.
(iv) Smoke.
(v) Moisture — mist, fog.
(vi) Blowing snow (3.10) or soil/sand (e.g. 'fen blows') (see 10.3.6.3).
(vii) Spray, at sea and at coastal sites (3.10).

In forecasting visibility, air-mass characteristics and history must be noted, particularly the possibility of importing pollution from continental sources and allowing for local sources. Fog forming in a polluted boundary layer is likely to have more, but smaller drops than fog forming in a cleaner air mass, hence the poorer visibility.

Much information on local visibility incidence (albeit prior to the 'Clean Air Act') is available in *Airfield Weather Diagrams and Characteristics* (extracts in Local Weather Manual for Southern England).

AWDC (1960)
Local Weather Manual for Southern England (1994)

3.2 Fog

Fog formation is complex; its occurrence is widely variable in space and time, forming under a wide range of meteorological circumstances. In all cases it forms as a result of air near the surface becoming saturated and being cooled below its dew point (see 10.3.3).

3.2.1 Classification
The basic fog classifications are: air mass fog((i) to (iii)) and frontal fog (iv):
(i) *Radiation fog* — loss of heat from the air, to ground cooled by nocturnal radiation, causes air to become saturated
(ii) *Advection fog* (warm and cold) — movement of air over a surface with different temperature and moisture regimes.
(iii) *Upslope fog* — air is forced, under stable conditions, up a land slope, resulting in adiabatic cooling and saturation.
(iv) *Frontal fog* — developing in frontal zones when precipitation from warmer air falls into very cold, stable air beneath, saturating the layer.

(*Mixing fog* — two unsaturated masses of air becoming saturated on mixing, is not a recognized mechanism for fog formation on the synoptic scale.)

On many occasions the development of fog is due to more than one factor.

Bader et al. (1995), Chapter 7
Perry & Symons (1991)
Thomas (1995)

3.3 Radiation fog

Radiation fog can occur at any time of the year but is most frequent in autumn and early winter; it is likely to be more widespread and persistent when *advection* brings an increase in vapour content just above the surface. Low-lying stations in eastern England experience an increased frequency of radiation fogs when there is a weak flow from the North Sea.

3.3.1 Physics of formation of radiation fog

(a) *Radiative cooling and dew deposition phase (Stage 1)*

(i) Under favourable conditions the absence of cloud permits strong radiative nocturnal cooling at the ground. Cooling is initially rapid; with (10 m) winds <7 kn a temperature inversion forms in the lowest levels near the ground.

(ii) Once the ground is at the dew point of the air, dew deposition onto the surface begins. This dries the air in contact with the ground and slows down surface cooling by the addition of latent heat.

(iii) Continuation of this process depends on the degree and effectiveness of the vertical mixing in the lowest layers; if sufficient mixing turbulence is present fresh supplies of moisture will be brought into contact with the ground and deposition of dew will continue. Hence the amount of water vapour in the atmosphere decreases (as shown by falling dew points) and air temperatures must fall further before fog can begin to form.

(b) *Initial formation (Stage 2)*

(i) Fog formation now depends on a delicate balance of several factors: principally the spreading upwards of radiative cooling by turbulence, and the drying out and warming of near-surface layers by dew deposition.

(ii) Initial formation often occurs when there is a lull in the 2 m surface wind to around 1 kn or less. This causes turbulence to cease and markedly reduces rate of dew deposition; with no significant air motion, any further radiative cooling in the lowest layers leads to supersaturation of those layers and hence condensation in the form of thin, shallow wisps of fog at a height of some 20 cm above the ground. (Note that there may not be a 2 m anemometer; under fog formation conditions it will not be possible to deduce the 2 m wind from a 10 m reading.)

(c) *Mature stage.*

With sky visible (Stage 3(i))

(i) Radiative cooling continues from the ground and the inversion base remains close to the surface.

(ii) As the screen temperature falls, fog gradually deepens. The upward heat flux decreases with time but may still be sufficient to halt the fall in surface temperature.

With sky obscured (Stage 3(ii))

(i) After a few hours the fog may be deep enough (65–165 ft; 20–50 m) to obscure the sky. The fog top now becomes the radiating surface, radiative cooling at the ground ceases and the surface temperature may start to rise, if the upward soil heat flux is large enough. The resulting heating at the ground and cooling aloft gives rise to convective overturning. In fact if the fog is deep enough to have an effective emissivity of ≈ 1, then cooling from the fog top can be sufficient alone to cause overturning.

(ii) This raises the inversion away from the surface as an SALR profile develops. Continued radiative cooling from the fog top causes the inversion to rise further and fog to continue to deepen.

(iii) Maximum inversion height of 230–750 ft (70–230 m) is generally reached in 2–3 hours after lifting from surface, with fog top about 80 ft (25 m) above base of inversion; there is a marked wind shear at the fog top.

The above stages are represented diagrammatically in **Figs 3.1(a)–(d)**.

Fig. 3.2 shows the conditions observed in a fog at Bedford in November 1983, which passed from its formative to its mature stage at about 01 UTC. Large-amplitude variations in fog top may be superimposed on the general upward growth.

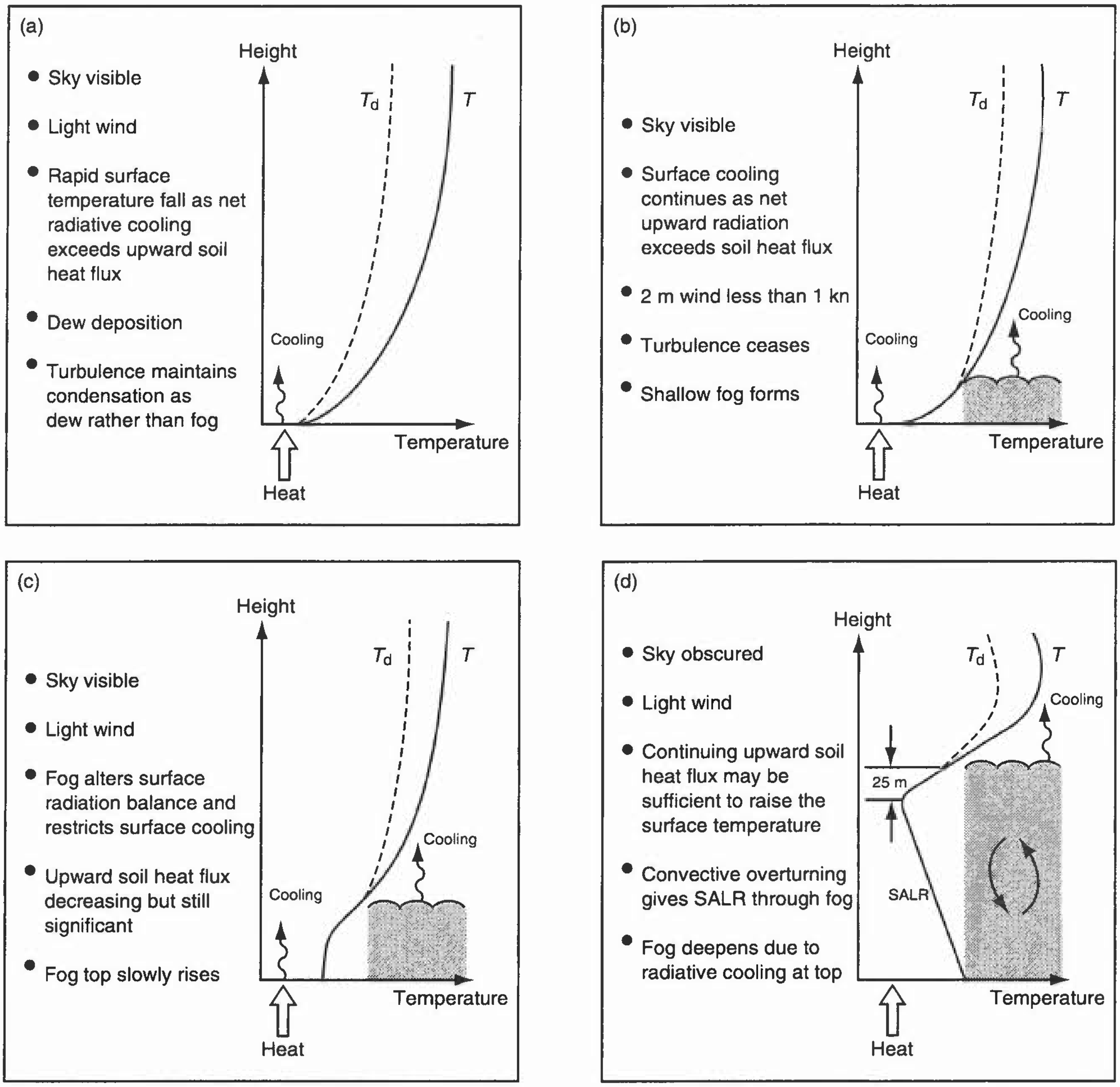
Bradbury (1989)	**Findlater (1985)**
Brown (1987)	**Roach (1994, 1995)**

3.3.2 Conditions observed during radiation fog formation

Factors favourable for formation:

(i) Clear skies or just thin, high cloud.
(ii) Moist air in the lowest 100 m or so.
(iii) Moist ground (e.g. after rain, or over marshes).
(iv) Favourable local topography.
(v) Slack pressure gradient, allowing the surface wind (preferably measured at 2 m) to decrease to near calm.

The frequency of geostrophic wind speeds (V_g) at Cardington during periods of fog were found to be as shown in **Tables 3.1 and 3.2**).



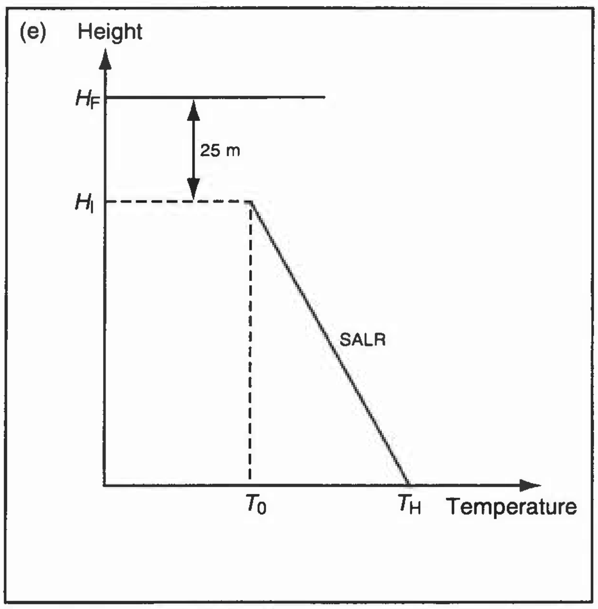


Figure 3.1(a–d). Schematic representations of (a) Stage 1 — radiative cooling and dew deposition, (b) Stage 2 — initial formation, (c) Stage 3(i) — development of a mature fog (sky visible), and (d) Stage 3(ii) development of mature fog (sky obscured).

Figure 3.1(e). Schematic daigram to estimate the height of the inversion base during sky obscured/fog deepening phase (Stage 3(ii) — Fig. 3.1(d)).

H_I Height of inversion base
H_F Height of fog top (H_I + 25 m)
T_0 Screen temperature when sky became obscured
T_H Screen temperature at time H after sky became obscured

See 3.3.2.1.

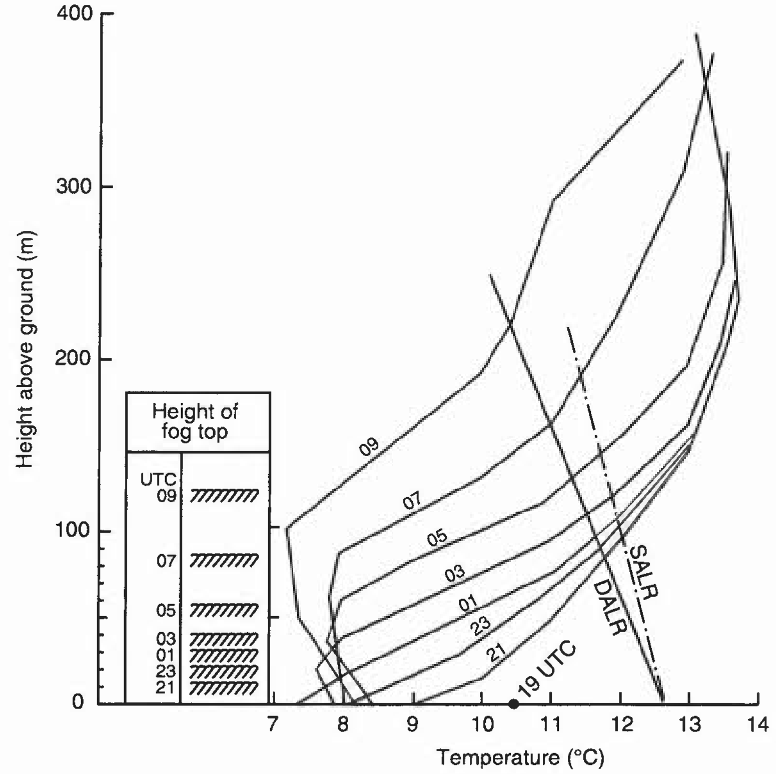


Figure 3.2(a). Temperature–height diagram showing the temperature structure and fog top at Bedford at various times during the night of 9/10 November 1983.

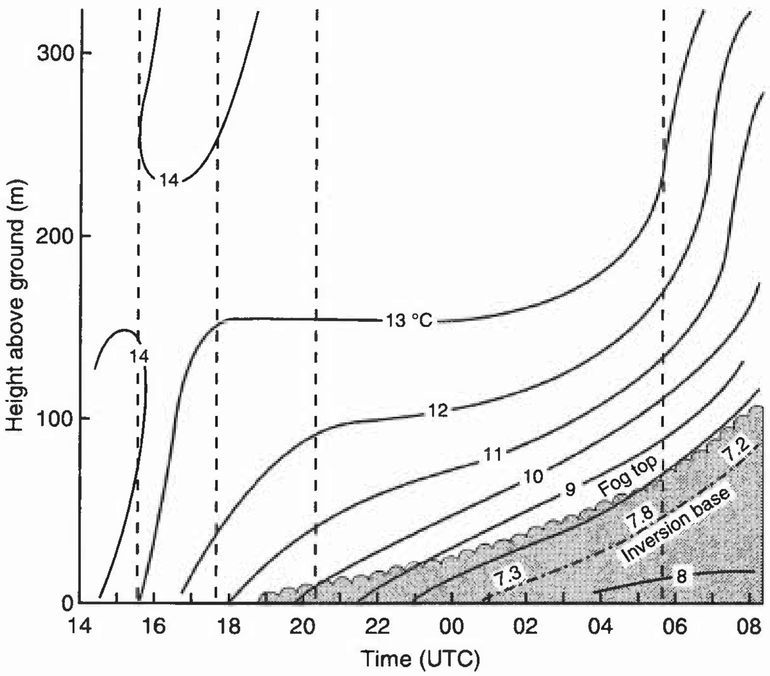


Figure 3.2(b) Time-section of the vertical temperature profile and fog depth at Bedford on the night of 9/10 November 1983. Fog formed at about 1830 UTC. The change from 'sky visible' to 'sky obscured' occurred just before 0100 UTC, which was also the time of the minimum surface temperature. The fog-top temperature from this time on remained steady, with the coldest temperature just below the fog top.

Table 3.1. Percentage frequency of V_g in various speed ranges

	knots										
	0–2	3–5	6–8	9–11	12–14	15–17	18–20	21–23	24–26	27–29	30+
A	7.1	16.7	23.8	23.0	18.3	10.3	0.8	—			
B	6.7	16.0	18.0	20.0	16.3	11.3	6.7	2.5	1.5	0.5	0.7

A = when fog first formed, B = while fog persisted

The wind speed at 10 m during the same fogs had the following frequency distribution:

Table 3.2. Percentage frequencies of surface wind in various speed ranges

knots	0–2	3–5	6–8	9–11	12–14	15–17
%	58.5	28.6	10.5	2.1	0.3	—

The stronger winds occurred when North Sea stratus spread inland replacing local radiation fog with widespread advection fog.

(i) Recent studies have shown that the wind speed at 1–2 m is more important than the 10 m wind. Fog forms when the 2 m wind falls below 2 kn. The wind at 10 m may be significantly stronger than the 2 m wind on radiation nights.
(ii) Synoptic studies demonstrate the preferred occurrence of radiation fog to be with anticyclonic pressure patterns, particularly on the western side (to where, generally, air of higher RH is moving).
(iii) Only about one fog in seven occurs in a cyclonic situation. Within these pressure patterns there is a greater tendency for fog to form in those sectors where there is a southerly component to the geostrophic wind (East Anglia/Lincolnshire) by a factor of about 3 to 1.

Findlater (1985)

3.3.2.1 Height of fog top (Findlater's method)

(i) During the period of fog deepening, and with minimum advective effects, the temperature at the base of the inversion was observed by Findlater to remain constant to within ±0.5 °C while the screen temperature rose slightly. In many cases, however, screen temperatures continue to fall, in which case the following technique is not appropriate.
(ii) Findlater's technique (based on only a small number of cases) is to draw the SALR temperature profile from the known screen temperature T_t at time t after sky obscuration to intersect an assumed temperature at the inversion base (i.e. the screen temperature when sky became obscured). The height of the inversion base, H_I, may then be estimated and fog top is taken as: H_I + 25 m (**Fig. 3.1(e)**) (this only applies while fog is still deepening and if advective effects are minimal).

Findlater (1985)

3.3.2.2 Factors modifying fog formation

Four specific factors which complicate and modify the radiative progress of fog formation are:

(a) *Advective effects*:
   (i) of fog formed upwind;
   (ii) of air with different temperature/moisture characteristics at low level;
   (iii) of sea fog inland before locally produced fog has formed;
   (iv) due to upslope motion (3.5).
(b) *Sunrise*: turbulence and added moisture due to the evaporation from the surface often give a sudden deepening soon after sunrise. Sudden fog formation may result if turbulence mixes cold surface air and warmer aloft.
(c) *Fog aloft*: very low stratus may be found with moderate winds; once formed continued cooling and turbulent mixing act to lower the base to ground level. The fog will only persist if winds become light.
(d) *Fog forming in smoky boundary layers* will have very poor visibility; formation of fog is less rapid in such cases possibly due to the reduction of the rate of radiative cooling of the ground and the distribution of condensing water among a large number of nuclei.

Findlater (1985)

3.3.3 Forecasting the formation of radiation fog

3.3.3.1 Calculation of fog point (Saunders' method)

The method takes account of moisture throughout the cooling layer; it is based on Mk. IIb radiosonde profiles; Mk. III T_d data may differ.

(i) Select a representative upper-air sounding and find the condensation level from the maximum temperature and the dew point at that time, using Normand's theorem (2.1.1).
(ii) Find the humidity mixing ratio at the condensation level and read off the temperature where the humidity mixing ratio line cuts the surface isobar. This is the expected fog-point temperature.

This procedure needs modification to allow for different types of sounding (see **Fig. 3.3**):

(a) *Type I* has a constant dew-point lapse rate except near the ground where the surface dew-point lies on, or to the right of, a downward extension of the upper dew-point curve. T is the maximum temperature and T_d is the surface dew-point. If there is a superadiabatic, use the value T_c instead of T to eliminate the superadiabatic section. The pecked lines through T_c and the dew point T_d meet at the condensation level A. The humidity mixing ratio at this level is at B and the fog point is at C.

(b) In *Type II* the dew-point lapse rate increases aloft. Point B is found by extrapolating the lower part of the dew-point curve above the point at which the lapse rate increases.

(c) In *Type III* the surface dew point lies to the left of the downward extension of the upper dew-point curve, two possibilities are illustrated:

  (i) If the temperature lapse in the lowest layers is less than a dry adiabatic, the construction follows the basic principles as for *Type I.*

  (ii) If the temperature lapse rate in the lowest layer is equal to or greater than a dry adiabat, then no Normand construction is drawn and the fog point is taken to equal the dew-point.

Normally, if the boundary layer is mixed at the time of the midday sounding, extending the mean mixing-ratio through the boundary layer to the surface will give an acceptable fog point. As a rough guide, the afternoon dew point at screen level minus 2 °C gives a good first guess for $T_f > -2$ °C.

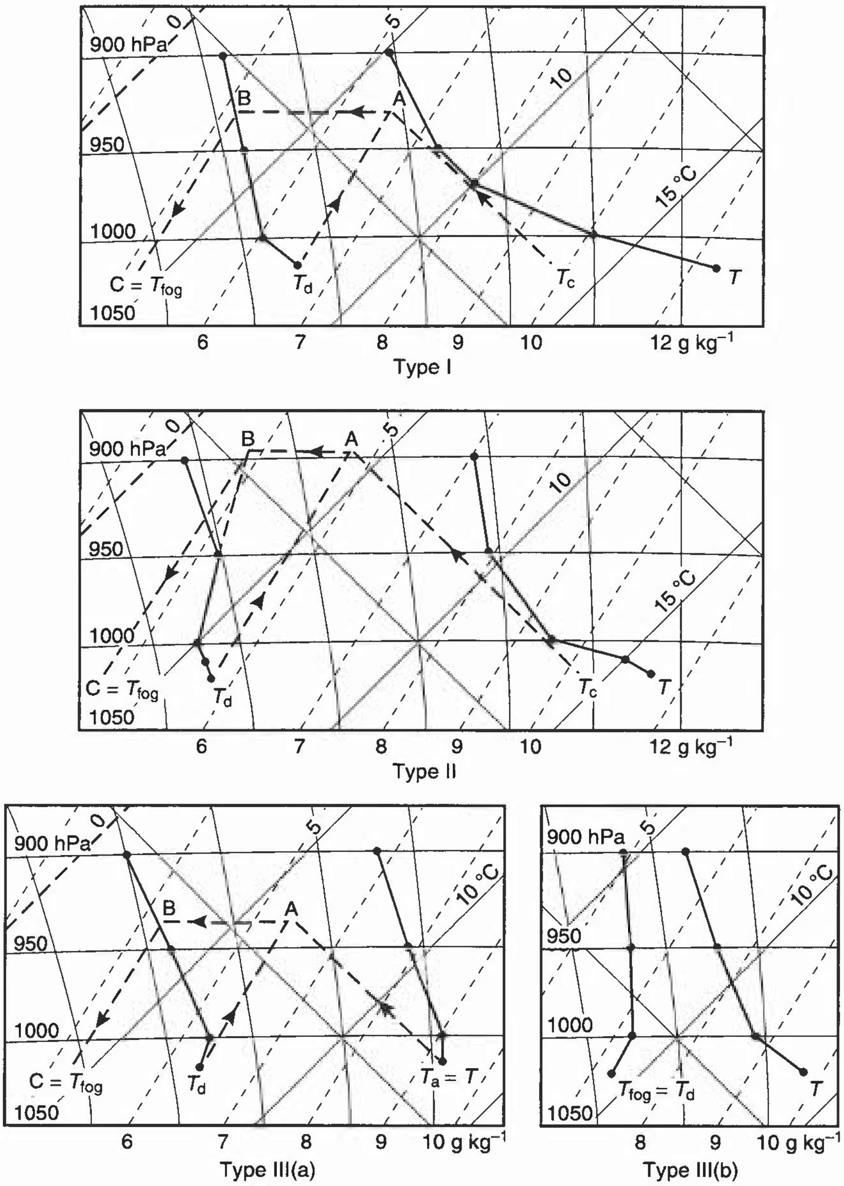


Figure 3.3. Estimation of fog-point (Saunders' method) with adjustments to midday soundings under various conditions. See text for explanation of types.

Notes:

(i) If a subsidence inversion has brought dry air down to within 30 hPa of the ground, use the dew point (T_d) as the fog-point.

(ii) If rain falls during the afternoon leaving the ground wet, the actual fog point may be higher than the calculated value.

(iii) If a sea-breeze reaches the area later in the day, the fog point may be much higher than calculated; use the coastal dew point.

(iv) The dew-point temperature 60 hPa above the surface nearly always gives a fog-point value close to Saunders' value, exception being *Type IIIb* when the surface dew-point is used.

(v) If the calculated temperature is ≤0 °C, then the actual fog point may well be lower (due to deposition by hoar frost).

Saunders (1950)

3.3.3.2 Calculation of fog-point (Craddock and Pritchard's method)

If T_f is the fog point, T_{12} is the screen temperature at 1200 UTC, and T_{d12} is the dew point at 1200 UTC, then

$$T_f = 0.044\,(T_{12}) + 0.844\,(T_{d12}) - 0.55 + A = Y + A.$$

Values of A and Y can be obtained from **Table 3.3**.

Table 3.3(a). Values of Y (°C) corresponding to the observed values of T_{12} and T_{d12}

T_{d12} \ T_{12}	30	25	20	15	10	5	0	−5	−10
20	17.7	17.4	17.2						
18	16.0	15.7	15.5						
16	14.3	14.1	13.8						
14	12.6	12.4	12.1	11.9					
12	10.9	10.7	10.5	10.2					
10	9.2	9.0	8.8	8.6	8.3				
8	7.5	7.3	7.1	6.9	6.6				
6	5.8	5.6	5.4	5.2	5.0				
4	4.1	3.9	3.7	3.5	3.3	3.0			
2	2.5	2.2	2.0	1.8	1.6	1.4			
0	0.8	0.6	0.3	0.1	−0.1	−0.3	−0.6		
−2	−0.9	−1.1	−1.4	−1.6	−1.8	−2.0	−2.2		
−4	−2.6	−2.8	−3.0	−3.3	−3.5	−3.7	−3.9		
−6	−4.3	−4.5	−4.7	−5.0	−5.2	−5.4	−5.6	−5.8	
−8	−6.0	−6.2	−6.4	−6.6	−6.9	−7.1	−7.3	−7.5	
−10	−7.7	−7.9	−8.1	−8.3	−8.6	−8.8	−9.0	−9.2	−9.4

The number A (°C) is an adjustment which depends upon the forecast cloud amount and geostrophic wind speed, as tabulated below.

Table 3.3(b).

*Mean cloud amount (oktas)	*Mean geostrophic wind speed (kn) 0–12	13–25
0–2	0.0	−1.5
2–4	0.0	0.0
4–6	+1.0	+0.5
6–8	+1.5	+0.5

*Mean of forecast values for 1800, 0000 and 0600 UTC.

Notes:

(a) The equation for T_f was derived from the combined data for 13 widely separated stations in England. There was considerable variation from station to station in their proximity to major smoke sources.

(b) Account was not taken of variations in atmospheric pollution so that, in effect, an average degree of pollution is assumed in using this technique (in contrast to Saunders' method which refers mainly to fog in clean air).

(c) If the minimum temperature is predicted using Craddock and Pritchard's method it is suggested that:

  (i) if T_f is 1 °C or more above T_{min}, forecast fog;

  (ii) if T_f is 0.5 °C above to 1.5 °C below T_{min}, forecast a risk of fog;

  (iii) if T_f is 2 °C or more below T_{min}, do not forecast fog.

(d) When forecasting for a region, rather than a specific airfield, allow a larger safety margin, since there is always more low-level moisture present near streams and in lush valleys than over flat airfields with trimmed grass.

Craddock & Pritchard (1951)

Other fog prediction techniques (Banks, Swinbank) are given in HWF; a 'probability predictor' diagram for Waddington, while not being statistically rigorous has operational credibility.

HWF (1975) Chapter 20.7

3.3.3.3 Inferred fog prediction from T_f − T_{min}

Table 3.4. Fog prediction summary

$T_f - T_{min}$ (°C)	Inferred fog prediction
≥+1	widespread fog expected
=0.5	fog probable late in the night
=0	patchy fog likely by dawn
−0.5 to −1.5	fog patches possible in fog-prone areas
≤−2	fog not expected

If T_f is significantly greater than T_{min} the time of formation can be predicted from the night cooling curve.

Craddock& Pritchard (1951)

3.3.3.4 The fog point in relation to the 850 hPa wet-bulb potential temperature

(i) The 850 hPa wet-bulb potential temperature is useful as a means of identifying air masses.

(ii) The probability of fog occurring increases as the temperature difference of the surface temperature below the 850 hPa WBPT increases, see **Fig. 3.4.**

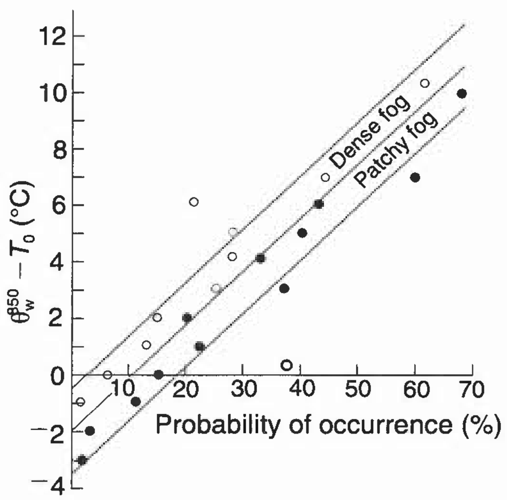


Figure 3.4. The probability of fog corresponding to a given depression of the surface temperature (T_0) below the 850 hPa wet-bulb potential temperature (θ_w^{850}).

3.3.3.5 Summary of forecasting procedures for fog formation

(i) Forecast the synoptic situation overnight
Air-mass characteristics
Cloud type and amount
Pressure pattern, gradient and surface winds

(ii) Consider history
Did fog form last night in the same air mass?
If so, was it widespread or patchy?
How will conditions differ tonight?
— Warmer or colder?
— Moister or drier?
— Cloudier or clearer?
— More or less windy?

(iii) Consider local factors
Look at local weather diagrams etc.
Topography, smoke sources, etc.

(iv) Estimate the fog point
Craddock–Pritchard
Saunders' technique
Any others available?

(v) Will the fog point be reached?
Construct night cooling curve
If $T_f > T_{min}$, what time will fog form?

(vi) Once the fog forms, what will the visibility be?
Large variations over short distances
Visibility often falls rapidly once fog forms
Usually less than 200 m in clean air
— Decrease in visibility may be slower in smoky air
If temperature <0 °C, 200 to 1000 m more common

(vii) Keep the forecast under review
Synoptic developments
Observations at your station
Observations from other stations nearby
Satellite imagery
Amend the forecast if necessary

The potential of a given location for thick radiation fog (visibilities <200 m) can be assessed by a Fog Potential Index.

Meteorological Office (1985)

3.3.4 Forecasting the clearance of radiation fog

Four main mechanisms for dispersing radiation fog:
(i) Solar radiation.
(ii) Advection of cloud cover over the fog top.
(iii) Increasing gradient wind.
(iv) Advection of drier air.

3.3.4.1 Fog clearance by insolation

Satellite imagery shows that extensive areas of radiation fog and stratus dissipate from their outer edges inward due to mixing generated by insolation (this can be a more-efficient mechanism than the absorption of solar radiation by fog and underlying surfaces). The time of clearance depends on the thickness of the fog, the location of the site relative to the fog boundaries and the insolation available at a particular latitude and time of year; hill/valley circulations complicate matters for valley fog. Estimates of fog duration may be made from fog brightness in VIS imagery.

Bader et al. (1995), Chapter 7
Gurka (1986)

The fog depth must be known or estimated, together with a fog-clearance temperature.

(a) *Estimation of fog top.*

(i) Visual estimation: if the sky is visible, the fog depth is probably about 5 hPa in dense fog, and 10 hPa in thin fog.

(ii) If the sky is obscured at dawn, and no local mini-sonde ascent is available, the most representative midnight sounding should be modified as in (b) below to allow for changes between midnight and the time of minimum temperature (note height of fog top is 25 m above top of inversion).

(b) *Modifying the temperature profile on a tephigram.*

Heffer's estimate is probably appropriate:

(i) If the nose has already formed on the temperature ascent curve, the level is raised by 5 hPa and the temperature decreased by 1.5 °C; this point is joined to the night-minimum temperature by a straight line on the tephigram. The point where this line and the downward extension of the dew-point curve intersect (O in **Fig. 3.5(a)**) represents the fog top.

(ii) If a nose has not yet formed on the midnight ascent, the point 35 hPa above the ground is joined to the night minimum surface temperature and the fog top at dawn estimated as before (O in **Fig. 3.5(b)**).

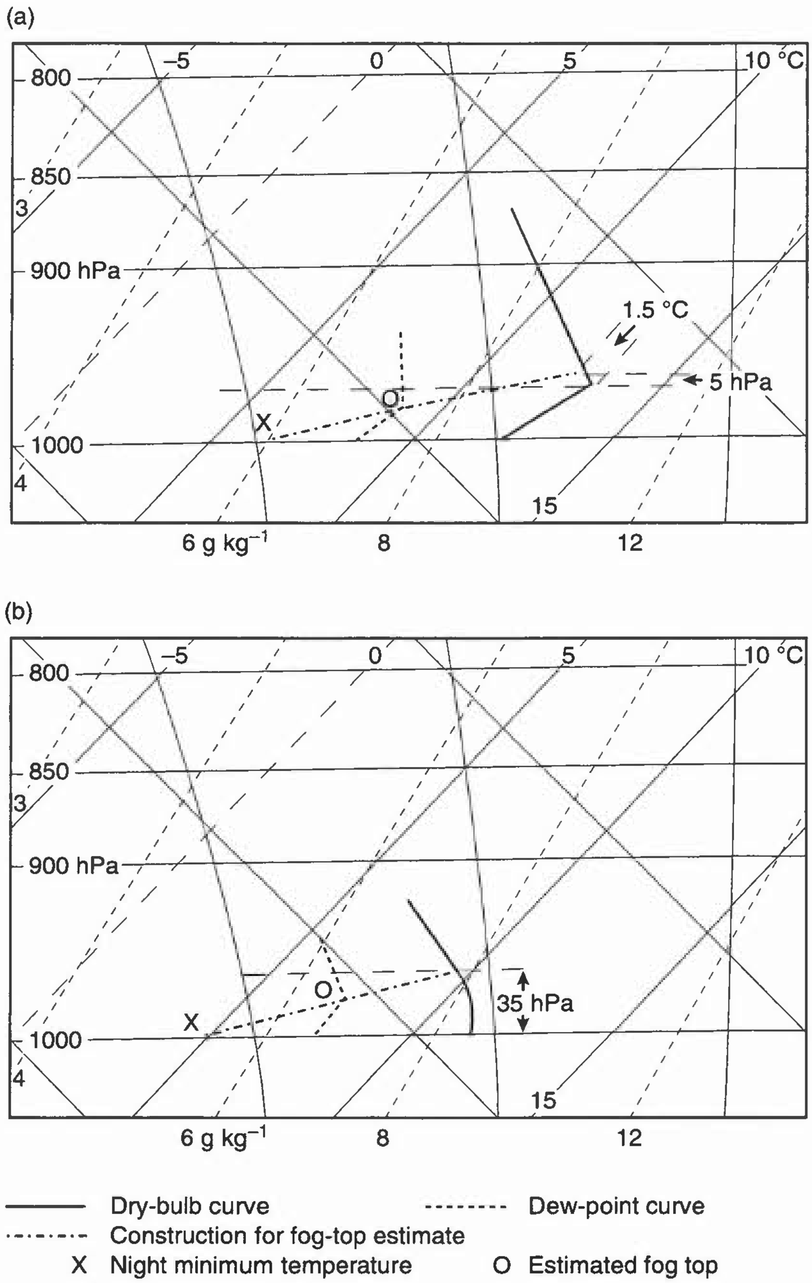


Figure 3.5. Construction for fog-top estimate from a midnight BALTHUM when (a) the inversion nose has already formed, and (b) the inversion nose has not yet formed.

(c) *Estimation of fog-clearance temperature (Jefferson).*
   (i) From the fog-top estimate, the clearance temperature is estimated on the assumption of a dry adiabat from the surface to fog top. Temperature rise is estimated from 2.3 (**Fig. 2.5**).
   (ii) The surface temperature thus obtained represents the probable value above which the fog will lift to very low stratus.
   (iii) The fog or low stratus should disperse entirely when the condensation level reaches the level of the inversion base (fog top will rise as fog lifts into stratus).

Caution: During winter when the sun is at a very low angle, the fog top will continue to radiate after sunrise. The fog will increase in depth for some time before the absorption of insolation is effective in starting the clearance process.

Barthram (1964) **Heffer (1965)**
Brown (1987) **Jefferson (1950)**
Findlater (1985) **Kennington (1961)**

(d) *A nomogram for forecasting clearance of fog by insolation*
Fig. 3.6 is a nomogram, due to Barthram, for predicting the time of fog clearance due to insolation, where:
T_1 = surface temperature at dawn
T_2 = fog-clearance temperature
d = depth of fog at dawn in hectopascals.

To use the nomogram:
(i) Enter the fog-depth on the left-hand diagram and move to the right to meet the vertical from the value ($T_2 - T_1$). From this point move downwards and to the right following the curved lines representing Q to the right-hand edge of the diagram.
(ii) Move horizontally across to the middle diagram and then as far as the vertical from the value $T_2 + T_1$. Then follow the curves for fQ to the right-hand edge of the diagram.
(iii) From this point move horizontally across to the right-hand diagram to meet one of the curves marked with dates. From this curve go down to the baseline where the time of clearance is marked.

If the fog is thin (visibility more than 600 m or depth <20 hPa) the insolation needed is reduced by one third. Take the value on the left-hand margin and follow one of the diagonal pecked lines down to the inner (thin fog) scale and continue as before.

Barthram (1964)

3.3.4.2 Fog clearance without insolation

(a) *Fog clearance following the spread of cloud*
   (i) The arrival of a cloud sheet over water fog often leads to the most rapid and efficient clearance of the fog because it stops, or reverses, the continual radiative cooling of the fog.
   (ii) The lower the cloud sheet the more effective it is in clearing the fog, provided the ground is not frozen.
   (iii) Heat flux from the soil, causing warming of the air by weak convective motions, may lift the fog into low stratus before its complete dispersal.
   (iv) The time taken for fog to clear decreases with higher temperatures, as is shown by the following observations (**Table 3.5**) made at Exeter Airport (south-west England).

Table 3.5.

	Initial grass temp. (°C)					
	<0	0–2	3–5	6–8	9–10	11–13
Average time (hours) for fog to clear after arrival of cloud sheet	3.1	2.2	1.1	1.5	0.9	0.5
Number of cases	10	10	5	10	5	3

Saunders (1957, 1960)

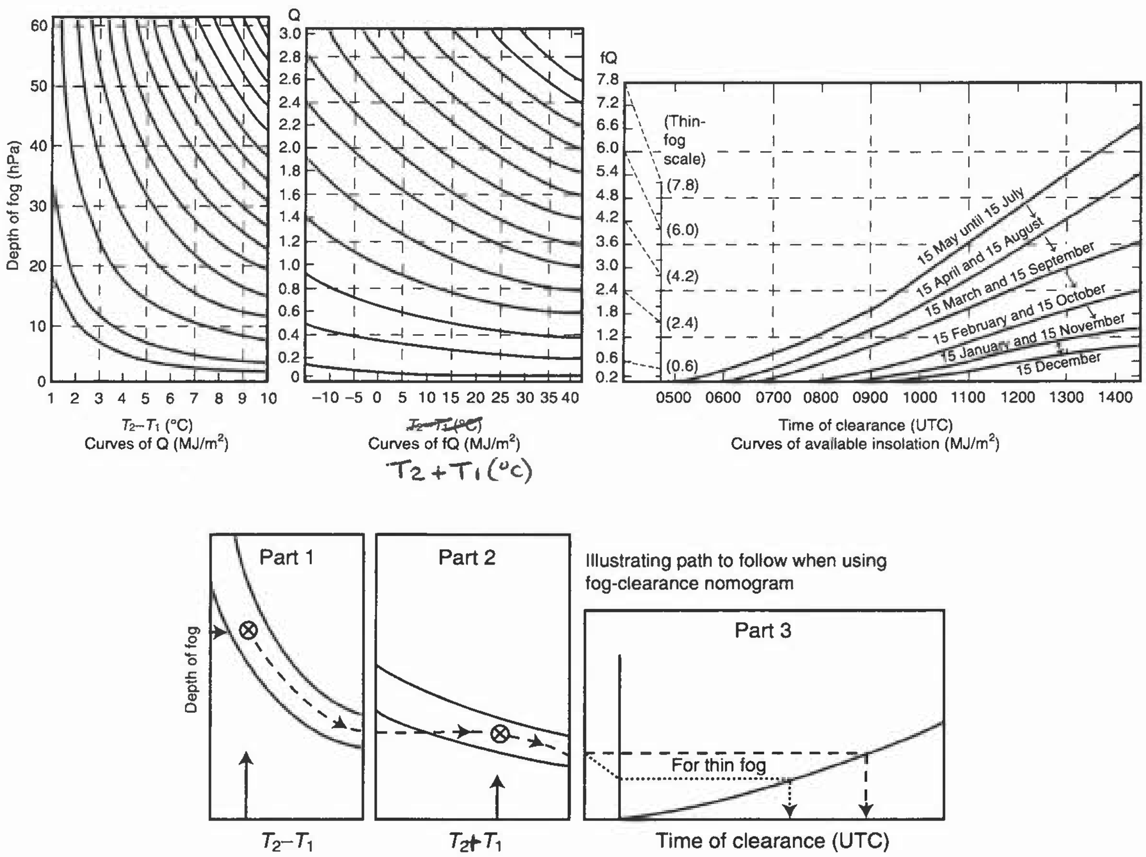


Figure 3.6. Nomogram for forecasting the clearance of radiation fog by insolation.

(b) *Fog clearance due to increase of gradient wind*

Although fog does not normally form unless the wind falls very light, its dispersal may be delayed until the winds aloft are quite strong. Mature fog has a well developed inversion capping it; through this mixed layer, a strong wind shear can be maintained. The more intense the inversion, the greater must be the wind above in order to produce turbulent mixing and dispersal of fog.

As a guide, the geostrophic wind speed required for fog to be dispersed by the increasing wind is:

(i) 15–20 kn over flat coastlands;
(ii) 20–25 kn inland;
(iii) 30–40 kn in deep valleys lying across the wind flow.

Even over the flat terrain at Cardington a peak geostrophic wind of 37 kn was estimated on one occasion before the fog cleared.

While increasing winds may lift fog into a layer of low stratus, localities downstream, especially where the ground has a gradual slope upwards, are liable to experience a delay in clearance, or the arrival of stratus if there was no fog initially.

HWF (1975) Chapter 20.8.5

(c) *Fog clearance due to dry-air advection*

(i) Local variations in air-mass surface characteristics and moisture content can create mesoscale patches of drier air which cannot be detected on synoptic-scale charts.
(ii) Unexpected nocturnal clearance of fog may occur when advection brings in drier air.
(iii) The gradual advection of progressively drier air can be effective in fog clearance.
(iv) The passage of a cold front can produce rapid fog clearance, by combining effects of dry air advection, increasing cloud cover and, often, increased wind speed.

3.3.4.3 *Persistent fogs*

Defining a persistent fog as one which forms one night and lasts all through the following day, not clearing until after dusk, then persistence occurs most commonly in eastern districts of England in winter.

(i) In December and January, one third of all fogs are persistent (according to this definition).
(ii) In November and February the proportion is about one sixth.
(iii) Care must be exercised in predicting whether overnight fogs between late November to early February will clear during the morning — the question is more 'if the fog lifts' than 'when'.
(iv) Most of these fogs are found to clear during the second night (44% by midnight and 80% by 0900 UTC on the next morning).
(v) Only 5–10% persist beyond 1700 UTC on the second day.
(vi) Radiation fog clearance in the summer half-year is infrequent and unlikely to persist after sunrise.
(vii) In autumn/winter it may persist overnight, although it tends to disperse next day.
(viii) Fog may thicken or, following a quiet, clear night, suddenly form soon after sunrise.
(ix) Most clearances are associated with the strengthening of winds aloft and/or an increase in low cloudiness, often (but not always) associated with a front within 200 km.
(x) Subtle indications of an impending clearance will include a slightly bigger pressure tendency or a change in wind drift at a particular station.

Note that, generally, the incidence of fog (and thick fog) has decreased markedly over the last 30 years as a result of the Clean Air Act legislation (1956).

Brown (1979)　　**HAM (1994) Fig. 59**
Kennington (1961)

3.3.4.4 *Summary of forecasting procedures for fog clearance*

(i) Study the synoptic situation
Will fog clear by any other mechanism?
— Increasing gradient wind?
— Advection of cloud cover over fog top?
— Advection of drier air?

(ii) Obtain a dawn temperature
Actual for your station if available
Otherwise forecast value

(iii) Choose a representative upper-air sounding
Latest time available (usually 2300 hrs)
Close to, or upwind of, station
Same air mass
Similar cloud/weather conditions
Beware of coastal influences

(iv) Modify the sounding
Sky visible or obscured?
Station dawn temperature

(v) Find fog clearance temperature and time
Kennington–Barthram technique
Other techniques available
Which one works best at your station?

(vi) Find stratus clearance temperature
— if fog is deep and widespread

(vii) Will local factors delay or advance clearance?
Upslope effects
Local wind circulations
Frost melt
Freezing fog can precipitate out

(viii) Keep the forecast under review
Synoptic developments
Observations at your station
Observations from other stations nearby
Satellite imagery
Amend the forecast if necessary

3.3.4.5 Large temperature falls associated with periods of weak advection on radiation nights

Sudden, large temperature falls (≥3.5 °C) have been recorded inland (at Lyneham, southern England) on radiation nights due to either:

(i) fog-free cold air being advected from adjacent low ground by an increase in wind, or:

(ii) fog forming in the same low area and deepening sufficiently to suddenly engulf the airfield.

On such nights it is clearly inadvisable to take a single station's observations as representative of an area.

Booth (1982)

3.4 Advection fog

3.4.1 Warm advection fog

Typical examples are sea fog and fog over very cold land. Thawing snow is commonly associated with advection fog over land.

(a) Factors favourable for *formation*:
- (i) Air-mass dew point greater than temperature of surface.
- (ii) Stable lapse rate, slight hydrolapse in lowest layers.
- (iii) Moderate winds (10–15 kn) — low stratus likely in winds greater than 15 kn.
- (iv) Suitable wind direction.

Note that: The advection of moist air also renders *radiation fogs* more frequent.

(b) Factors favourable for *clearance*:
- (i) Change of air mass — the most common and reliable means.
- (ii) Change of track of air mass to drier source.
- (iii) Over land — solar heating; radiation fog clearance techniques may be used. (Over snow the surface temperature may not rise above 0 °C).

HWF (1975), Chapter 20.9

3.4.1.1 Sea fog

There are four main sources of warm, moist air giving sea fog over coasts and waters around the British Isles:

(i) A south-westerly flow from the Atlantic west of the Iberian Peninsula; the air should have originated from, or spent some time over, warm waters south of 40° N. It is often associated with the warm sector of a frontal depression; it gives rise to widespread fog around southern and western coasts.

(ii) With high pressure to the west of the British Isles, mild air may circulate around northern Britain and come southwards as a north to north-east wind, bringing low stratus and fog ('haar') to exposed eastern coasts of Scotland and north-east England.

(iii) Warm continental air during spring and early summer in a south-easterly flow from the Mediterranean across Europe may become sufficiently moist and stable at low levels to affect east coast areas of the United Kingdom.

(iv) In summer very high dew-point air often accompanies thundery lows moving north from France, giving rise to sea fog in the English Channel and southern North Sea which may spread inland at night.

Bader et al. (1995), Chapter 7
HWF (1975), Chapter 20.9
Roach (1995)

3.4.1.2 Prediction of sea fog

Satellite imagery (VIS) is invaluable for locating the boundaries of existing areas of sea fog:

(i) The shape of the foggy areas may correspond to areas of particularly low sea temperature (for example along the coast of north-east England).

(ii) Other patterns are due to the incursion of tongues of much drier air, usually of continental origin.

For prediction:

(i) A detailed and up-to-date chart of sea surface temperatures is essential, combined with an analysis of the dew-point distribution over the sea and a prediction of the low-level air flow.

(ii) Wind speed in sea fog is most commonly 10–15 kn, but it is not unusual to find winds of 25 kn. In some cases the wind speed may reach gale force, though such observations are usually confined to ships in mid-Atlantic.

3.4.1.3 Advection of fog from land to sea

(i) Radiation fog carried out to sea by a light offshore wind may drift considerable distances over a slightly warmer sea before being dispersed.

(ii) Radiation from the fog top may effectively disperse heat supplied to the lower layers of the fog by warmer waters. This results in a mature fog type of profile.

(iii) Occasionally such fogs may be carried back to land by a sea-breeze.

3.4.1.4 Advection of fog from sea to land

(i) Sea fog is frequently carried inland over coastal districts by winds off the sea (during the summer months sea-breezes may be responsible for this).

(ii) Nocturnal penetration may be extensive over low-lying ground where the liability to fog will be increased by radiational cooling.

(iii) By day sea fog usually lifts to low stratus overland and 'burns off' with adequate insolation.

(iv) Such clearances start off well inland and gradually spread upwind towards the coast.

3.4.1.5 Advection fog over land

(i) For advection fog over land to be at all widespread or persistent it is generally necessary for the ground to be very cold — either frozen or snow covered. Particularly widespread and persistent fog may occur when warm air starts snow cover thawing.

(ii) In summer the fog tends to spread inland in the evening and burn back to the coast in the morning. Fog often returns to an inland station when the temperature falls to the value at which the fog cleared in the morning.

(iii) During the coldest months, when insolation is too weak to disperse mature fogs, any change in wind speed or direction needs to be watched for signs that persistent fog patches may be advected into previously clear zones. Such movements can occur even in the middle of the day when a deterioration is least expected.

3.4.2 Cold advection fog

Cold advection fog forms when cold dry, stable air flows over a much warmer water surface (warmer by 10–15 °C at least).

(i) The evaporating vapour immediately condenses again to form *steam fog*, with convective whirls and a top at a few metres.

(ii) In polar regions the same process gives rise to 'Arctic Sea Smoke'.

HWF (1975) Chapter 20.9

3.5 Upslope fog

3.5.1 Requirements

(i) flow of moist air over gently rising ground over a wide area **(Fig. 3.7)**;

(ii) a stable lapse rate in the lowest layers.

(Very stable air may be deflected around the edge of steeply rising ground if there are gaps in the escarpment.)

On radiation nights, weak moist advection may be combined with gentle upslope motion producing multiple conditions favourable for fog; the likelihood of formation of upslope fog may be determined from a representative tephigram — the height of the base being given by the lifting condensation level of the upwind air.

3.5.2 Hill fog

(i) Hill fog does not necessarily require upslope motion.

(ii) It occurs when the cloud base happens to be lower than the hill top **(Fig. 3.8(a))**.

(iii) Depending on the humidity profile below the general cloud base, upslope motion may generate cloud on hills at a lower level, either merging with the general cloud layer or as a separate layer well beneath it.

(iv) Radiation fog which has formed overnight in valleys may lift into low stratus during the morning, covering hills temporarily before dissipating **(Fig. 3.8(b))**.

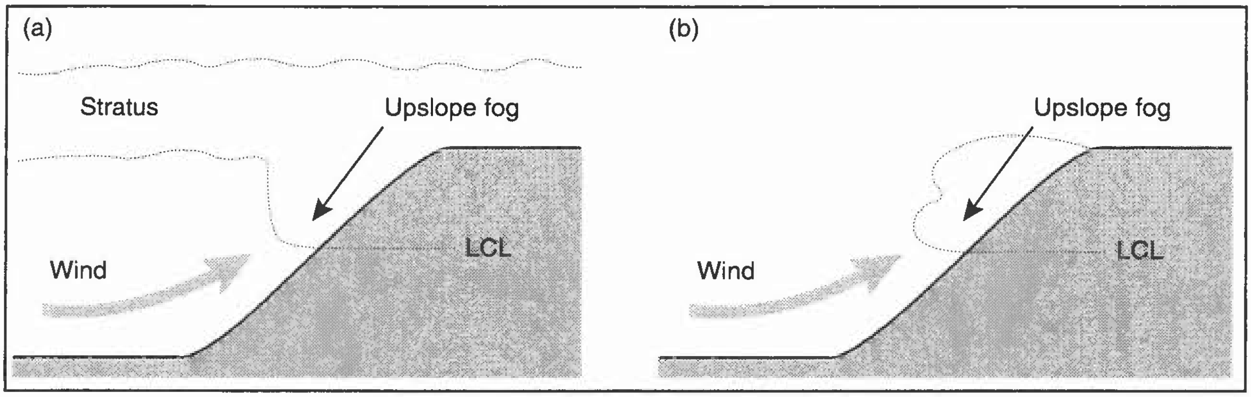


Figure 3.7. Upslope fog formation. (a) If turbulence is already causing stratus over flat ground, the effect of upslope motion will be to lower the cloud base on the windward side. (b) If turbulence is insufficient to form stratus, the extra uplift generated by upslope motion may cause patchy upslope fog to form. This can sometimes be the first sign of stratus forming more generally.

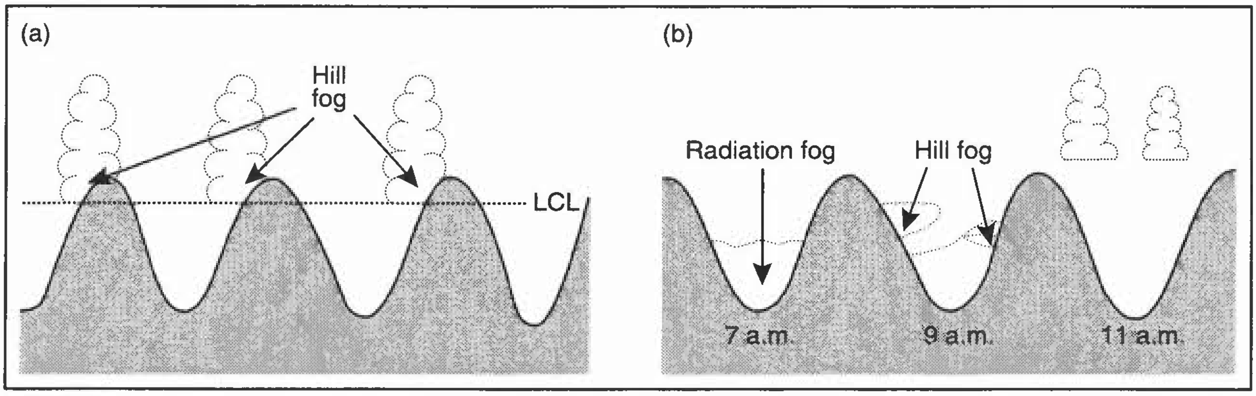


Figure 3.8. Hill fog formation. (a) Hill fog can occur in any air mass provided the hills are high enough, or the cloud base low enough. Here the base of the cumulus clouds is low enough to give hill fog patches. (b) Radiation fog which has formed overnight in valleys may lift into low stratus during the morning, covering hills temporarily before dissipating.

In the presence of strong winds the empirical relationship is suggested for the height of the cloud base in kilometres:

Cloud base (km) = dew-point depression (°C)/8;

thus in a southerly airstream ahead of a depression, with dew-point depression near a coast of <2 °C, cloud base may develop only 250 m asl.

Pedgley (1967)

3.6 Frontal fog

(i) Frontal rain falling into very cold stable air beneath eventually saturates the layer, frequently forming *pannus*, with fog confined to higher ground **(Fig. 3.9)**.
(ii) Occasionally winds may be light enough to give fog at low levels.
(iii) Visibility is often several hundred metres; forced ascent may reduce this.

3.6.1 Summary of factors favourable for frontal fog development:

(i) Ahead of active warm front or warm occlusion.
(ii) Very cold (or snow-covered) ground.
(iii) Large temperature contrast between cold and warm air masses.
(iv) Light surface winds.

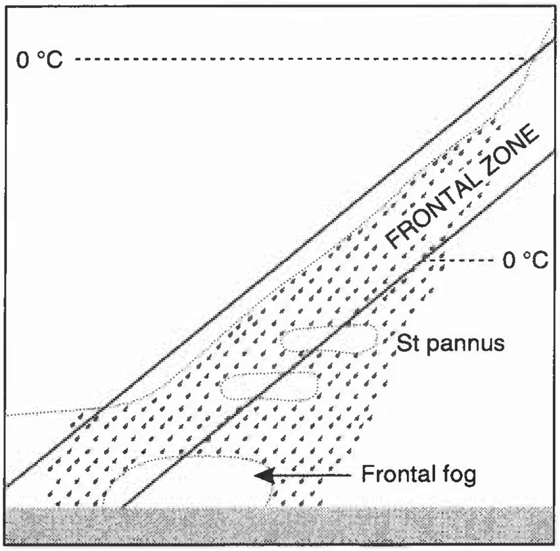


Figure 3.9. The formation of stratus pannus and frontal fog.

3.7 Convective activity above fog

(a) Although fog is normally associated with a stable air mass, the top of the inversion may be below the level of adjacent hills. During the day fog-free high ground may become warm enough to set off cumulus clouds.
(b) Beware of the development of thermally induced circulations in the low-level air flow; these may start the fog bank moving.
(c) Where large areas of low-lying fog persist throughout the day the heating of fog-free areas may result in the development of a pseudo sea-breeze effect during the afternoon. In winter, foggy air may be advected for many miles to produce sudden deteriorations in previously clear localities.
(d) In south and east England, wind speeds in excess of 15 kn have been observed as the fog front advanced.
(e) Fog formed in warm moist air advected over the sea may be overlain by a potentially unstable layer at higher levels. Convergence in the layers above the fog may set off vigorous convective activity and thunderstorms. In the early stages these may not affect the fog but the downrush of air associated with heavy storms will lead to localized clearances.

The following features may be used as a guide:
(i) Upper-level convective developments occur with a south-easterly flow at low levels and relatively high dew points near the surface.
(ii) Wet-bulb potential temperatures show a decrease with height in the middle levels.
(iii) The potential instability is realized if an upper low or trough in the contours at 500 and 300 hPa moves across the area.
(iv) This development will become apparent on infrared satellite imagery where the cold tops of instability cloud appear white above the grey of sea fog. Beware of situations of fog and thunder giving rise to 'anaprop' on the radar screen (10.6).

3.8 Guidance on the formation and detection of fog through imagery

Satellite imagery will help to determine several atmospheric conditions that can lead to fog formation:
(i) *Low-lying moisture* — prior to fog formation surface appears warmer under low-level moist air than where air is drier. Fog boundaries are generally more distinct than moisture boundaries.
(ii) *Cloud cover* — likely locations are areas with daytime cloud cover, or with a trajectory from cloud-covered areas, followed by clear skies at night.
(iii) *Snow cover* — snow cover can be a source of cooling that leads to fog formation.
(iv) *Precipitation* — if during the late afternoon or evening, followed by clearing skies will often lead to fog formation.

Extra information may be deduced from satellite multi-spectral channels.

Bader et al. (1995), Chapter 7

3.9 Haze

3.9.1 *Haze particles*

These may be produced by:

(i) Industrial pollution (of local and/or continental origin) and vehicle exhausts and may develop due to photolytic reagents; thus visibility may be substantially reduced in the 'urban plume' downwind of a city (2.11).

(ii) Smoke concentration near the ground will be less in stronger winds; it will only seriously affect visibility when atmosphere is stable at some level in the lowest 500 to 1000 m (1650 to 3300 ft).

(iii) Dust and fine sand raised by the wind during prolonged dry weather (e.g. 'fen blows').

(iv) Salt spray produced by rough seas and carried inland by strong winds.

3.9.2 *Haze occurrence*

(i) Haze is often concentrated where RH is high (e.g. under an inversion); hygroscopic particles can result in haze at relatively low RH. Injection of aerosols through the inversion by convective activity can lead to haze in stable layers well up in the troposphere.

(ii) The concentration of particulates will be modified by diurnal variations in temperature and RH as well as cloud cover (affecting the temperatures in the lowest layers) and precipitation (wash-out agent).

(iii) Urban haze may thicken in the afternoon due to photochemical reagents.

(iv) Haze is a problem for aircraft in low-level flight, especially viewing objects through a slant-path and particularly when looking towards the sun.

(v) Haze boundaries are often sharply defined, and may be associated with convergence lines (e.g. sea-breeze fronts).

(vi) The position of weak synoptic-scale fronts may also be located at times by haze boundaries.

AWDC (1960) **HAM (1994)**
Bradbury (1989) **Hänel (1976)**

3.9.3 *Depth of haze*

(i) With a well-marked stable layer near the ground, the haze top is often very clear-cut.

(ii) The tops of shallow cumulus generally extend just above the haze layer, **Fig. 3.10(a)**. If the air above is conditionally unstable, large cumulus or cumulonimbus may rise many thousands of feet above the haze top, which is usually at, or below, the level where the environment lapse rate decreases to less than the SALR.

(iii) Dispersing layer clouds may leave thin layers of elevated haze. These are rarely apparent from the ground but can be seen from aircraft when viewed horizontally during climb or descent.

Bradbury (1989)

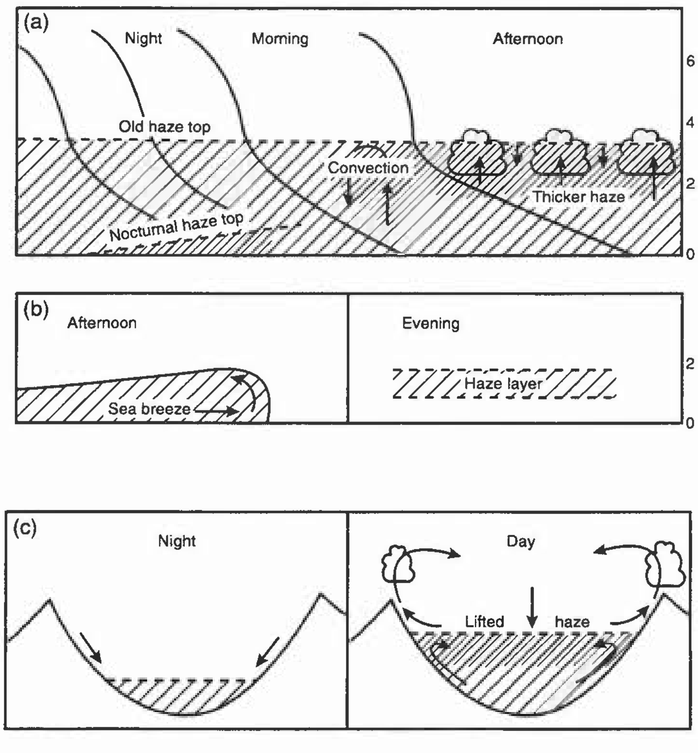


Figure 3.10. Idealized patterns of the diurnal variation of haze. (a) Under calm conditions with a stable layer at 3000–4000 ft. Curved full lines give an impression of the temperature profiles. (b) Association of haze with sea-breezes, which degenerate in the evening leaving an elevated haze layer. (c) Valley haze. Daytime convection over surrounding high ground generates subsidence over the valley which keeps the haze trapped in a restricted depth.

3.9.3.1 Airborne visibility
A forecaster should be aware of the effect on visual contact with the runway of slant visibility and be able to give guidance as to the likely height at which poor visibility may be encountered. Air-to-ground visibilities are made worse into the sun by reflection and scattering from the haze/fog top. Flight below an inversion may make target acquisition impossible, while above the inversion visibility may be much improved. (Visibility air-to-ground and air-to-air is also a function of column size, illumination, etc.)

HAM (1994) Chapter 9.2

3.9.4 Diurnal variation of haze
(i) At the surface, haze thickens during the night as RH increases and the nocturnal low-level inversion intensifies.
(ii) It thins out during the day when the RH decreases and the inversion is destroyed by solar heating.
(iii) Mesoscale weather systems exert important controls on the distribution of haze. The sea-breeze circulation, for example, with its diurnal rhythm, often very clearly distinguishes the well mixed, convective air inland from the much hazier, stable sea air (**Fig. 3.10(b)**).
(iv) Daytime convective mixing increases the haze depth and improves the horizontal visibility reported at the surface, but during the day the haze thickens at the inversion level where RH is highest. This effect is particularly marked when the inversion is low and haze is being channelled into the valleys intersecting areas of high ground. Valley haze (**Fig. 3.10(c)**) is due to subsidence generated by daytime convection over surrounding high ground.

HWF (1975), Chapter 20

3.9.5 Dispersal of haze
Continuous rain is effective in washing out most of the haze particles from the air. Showers, and even heavy thunderstorms, are much less effective.

3.9.6 Synoptic situations favourable for haze
(i) Haze is usually thickest in anticyclonic conditions when low-level winds are light. An increase in surface wind often leads to improved visibility, and a decrease in wind to poorer visibility at the surface.
(ii) Ahead of a warm front, when the clouds are increasing, the visibility from the air often deteriorates in the layer extending about 300 ft below the cloud base.
(iii) In the United Kingdom a high proportion of haze days are associated with winds from Europe (**Fig. 3.11**). Surface wind directions from 060–120° commonly bring the worst haze.

3.9.7 Visibility forecasting methods

3.9.7.1 The File method
Based upon data for five stations in southern England, the method is intended to be used 6–8 hours before the forecast time. Part I requires prediction of afternoon and evening wind speed and relative humidity. Forecast values are read from **Fig. 3.12.** In Part II a correction factor for those results is obtained by considering the surface chart and inspecting upwind humidity and wind speed and then applying these to the nomogram.
Example: actual visibility is 20 km, relative humidity about 60% and surface wind 10 kn. **Fig. 3.12(a)** gives a visibility of 13 km. The correction factor is then 20/13 = 1.5; this is applied to all subsequent forecasts. (A good forecast of advective changes to the airstream is necessary.) File recommends that in highly polluted airstreams visibility derived from the graphs be halved.

File (1985)

3.9.7.2 A (non-frontal) visibility forecasting method devised at Middle Wallop
(i) Construct a forecast night-cooling curve; find intercept of expected air-mass dew point with cooling curve.
(ii) Find out visibility at time of the maximum temperature (Vis_{max}); allow for areas of differing visibilities advecting across area.
(iii) Forecast overnight surface wind.

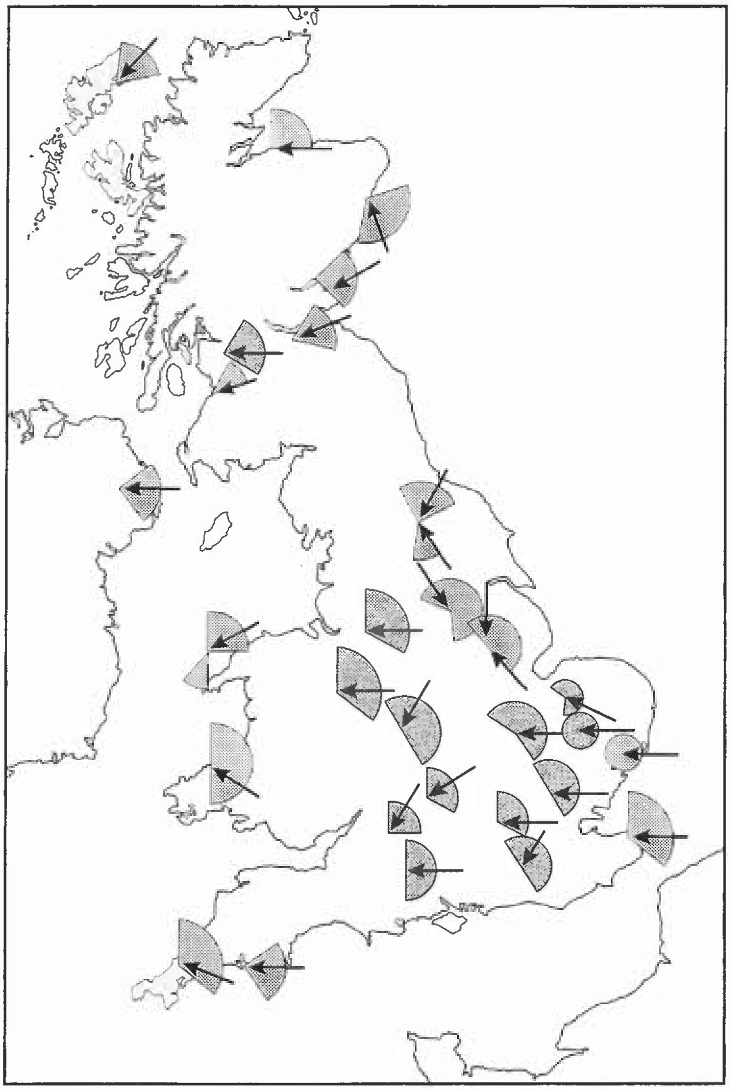


Figure 3.11. Surface wind direction and summer haze. Shaded sectors show the range of directions most often associated with visibilities in the range 1.8 to 9.9 km. Where variations are small the area shown is circular, and the arrows denote the worst directions. The radius of the sector arcs has no significance.

Then visibility, when temperature has fallen to air-mass dew point, depends on wind speed (**Table 3.6(a)**):

Table 3.6(a).

Wind (kn)	Visibility
0–5	⅓ of Vis_{max}
6–9	½ of Vis_{max}
≥10	¾ of Vis_{max}

Thereafter for each °C below the intercept temperature halve the visibility, noting time from curve.
For example, dew point at T_{max} is +4 °C, when visibility was 25 km; forecast wind speed: 3 kn.

Table 3.6(b).

At forecast temperature of 4 °C, forecast visibility is 8 km
At forecast temperature of 3 °C, forecast visibility is 4 km
At forecast temperature of 2 °C, forecast visibility is 2 km

A further proposition is that visibility at dusk is 80% of Vis_{max}.

Perry & Symons (1991)

3.9.7.3 Changes in horizontal visibility with height associated with relative-humidity changes

An untested, but theoretically sound, method for inferring the variation of visibility with height from a given surface observation is derived as follows:

(i) Aerosol is assumed well mixed, relative humidities (*RH*) >75% (0.75) and <98% (0.98) and with no local sources.

(ii) The relationship between the horizontal visibilities V_1 and V_2 at RH_1 and RH_2 is:

$$V_2 = V_1[(1 - RH_2)/1 - RH_1]^N$$

where $RH_2 > RH_1$ and N applies to a particular location, possibly varying with season, wind direction, etc. (German studies for an urban aerosol give $N = 0.7$; this may be ascertained for the local area by plotting log of visibility against log of $(1 - RH)$, when the slope will give N).

(iii) Using **Fig. 3.13**, plot starting surface visibility and associated RH.

(iv) Draw a line through this plot whose slope corresponds to the local N (shown here for values 0.6, 0.7 and 0.8; illustrated are other V', RH' relationships, where V at $RH = 0.8$ is assumed known, perhaps typifying local wind directions etc).

(v) Read off visibilities at other heights, corresponding to the (higher) RHs at these heights (derived from tephigram estimates).

Hänel (1976)
Reichert (1978)

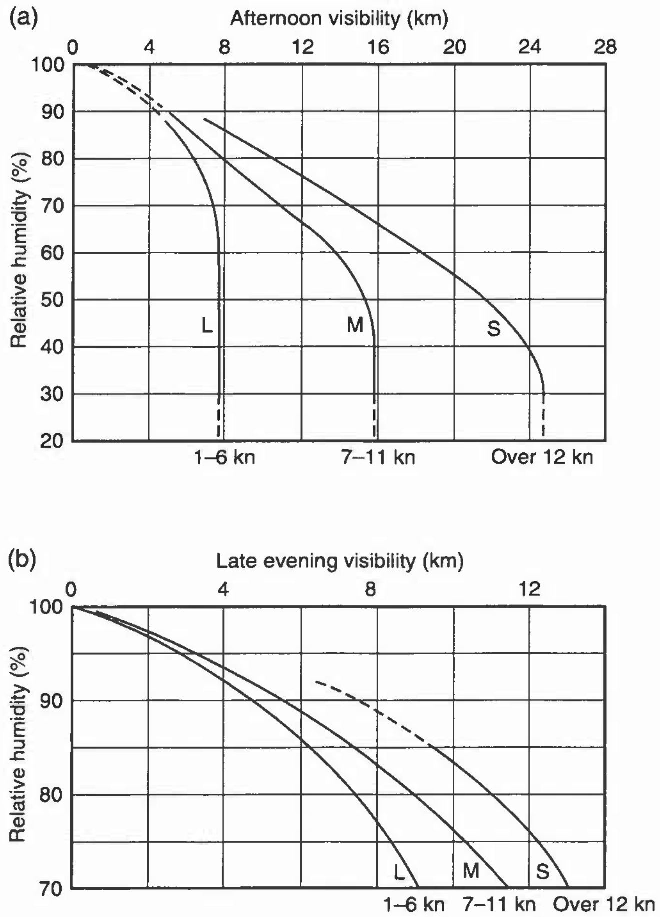


Figure 3.12. (a) Afternoon visibility against relative humidity for three classes of wind speed in 'easterly' synoptic situations. Median values are represented by L, M and S for light, moderate and strong wind speeds, respectively. The curves should be regarded as provisional for relative humidities above 87% and below 30%. (b) Late evening visibility against relative humidity for three classes of wind speed in 'easterly' synoptic situations. Median values are represented by L, M and S for light, moderate and strong wind speeds, respectively. The curve for strong winds should be regarded as provisional for relative humidities above 87% where data were sparse.

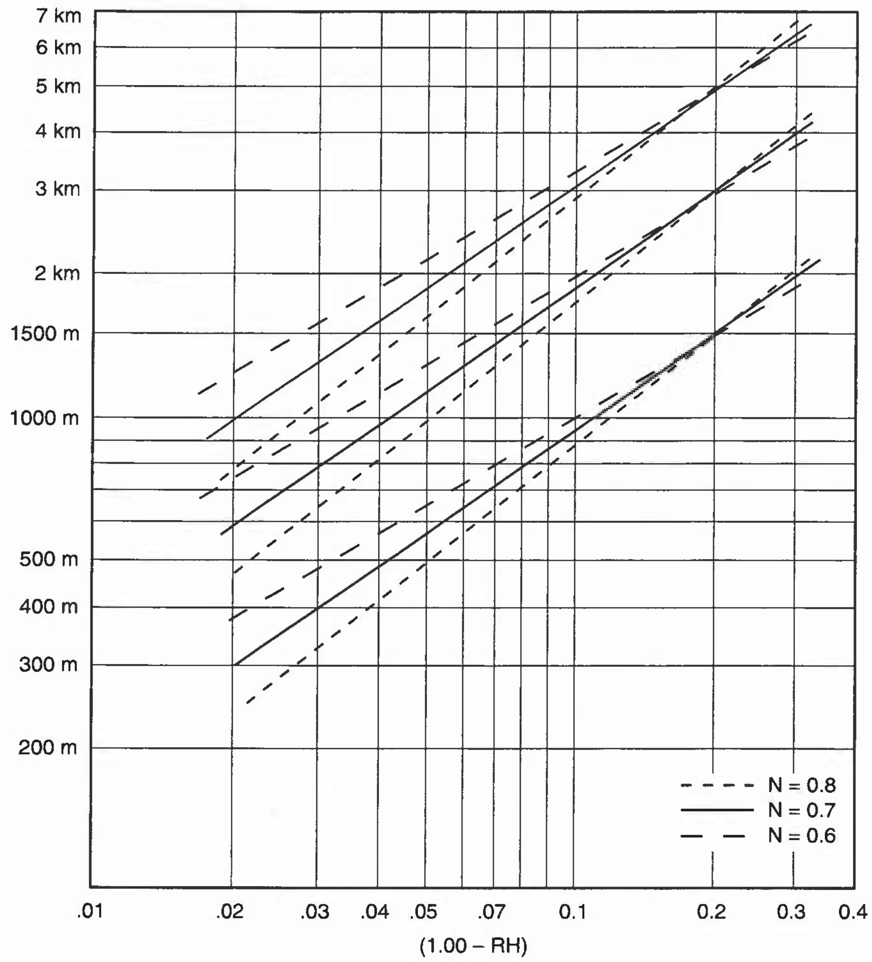


Figure 3.13. Diagram for estimating changes in horizontal visibility associated with changes in relative humidity.

3.10 Visibility in precipitation and spray

Rain — visibility is inversely proportional to total water content and number of raindrops; thus deterioration is greatest in heavy rain and in drizzle. **Fig. 3.14** summarizes experience in various countries; in moderate rain visibility is between 5 and 10 km, while heavy showers can reduce visibility to 1000 m, assuming no pre-existing pollution.

Snow — has a greater impact, visibility commonly falling below 1000 m even in moderate snow; in heavy snow this may fall to 200 m or less. Dry snowflakes result in visibilities only about one half of those illustrated in **Fig. 3.14** (wet snowflakes collapse to a smaller volume and become translucent). Blowing snow gives very low visibilities. It is most likely to occur when snow is dry and powdery.

In drizzle visibility ranges from about 3000 m to 500 m although simultaneous presence of fog droplets will still further reduce range.

In *low cloud* (hill fog) visibility is reduced to below 30 m.

In *spray* at sea and at coastal sites — winds >50 kn reduce visibility to <5 km
winds >70 kn reduce visibility to <1 km.

Jefferson (1961)

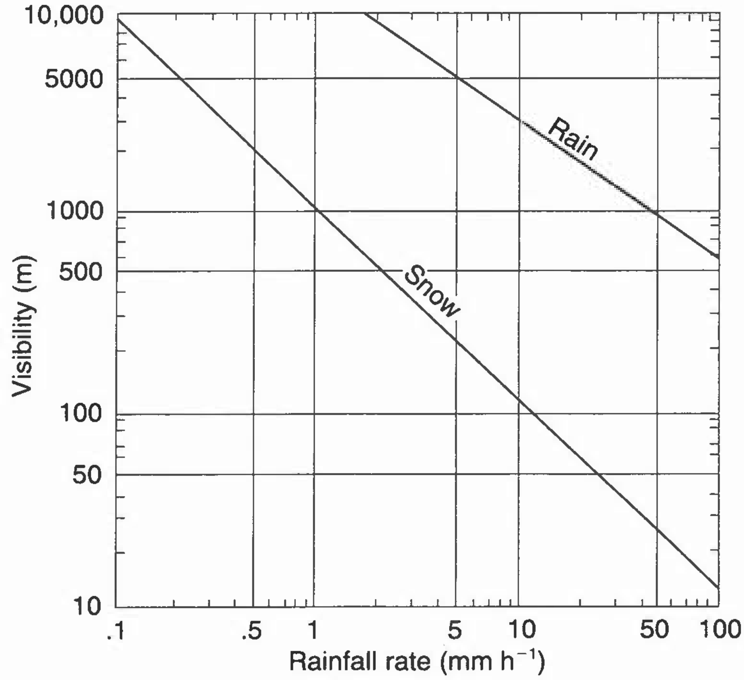


Figure 3.14. Visibility in rain and snow.

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