CHAPTER 5 — LAYER CLOUDS AND PRECIPITATION

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5.1 Layer cloud formation

Low stratus forms when air with a lapse rate less than the SALR is either cooled below its dew point or has extra moisture added by the evaporation of falling precipitation or by evaporation from wet surfaces, especially when snow or heavy frost begins to thaw.

Much information on the relation of stratus to local wind direction is available in *Aerodrome Weather Diagrams and Characteristics* (extracts in Local Weather Manuals).

AWDC (1960)
Mansfield (1988)

5.2 Large-scale ascent

5.2.1 Frontal cloud

Gentle ascent will often produce layer clouds throughout an extensive tropospheric depth; the following are forecasting techniques for short and longer term:

(a) Short term:
  - (i) Tephigram analysis — **Table 5.1** may give an indication of cloud structure; high cloud may obscure lower cloud on satellite imagery; a general sense of cloud structure may be inferred from radar rainfall.
  - (ii) Advection — apply gradient wind component normal to front; actual/forecast winds at cloud level; speed of warm/cold advection from hodograph.
  - (iii) Development — refer to pressure tendencies; frontal waves; upper-air development; local effects.

(b) Longer term — beyond 12 hours:
  - (i) Model output to position systems and indicate their likely activity.
  - (ii) Conceptual models (7.1) and local knowledge to estimate likely cloud structure.

Table 5.1. Cloud structure from a tephigram

The following guidelines should give a reasonable assessment of the likely cloud structure from dew-point depressions on a representative tephigram.

Dry bulb >0 °C	
$(T - T_d) \leqslant 1$ °C	8/8 layer
$(T - T_d)$ 1 to 5 °C	Thin layers
$(T - T_d) > 5$ °C	No cloud
Dry bulb <0 °C	
$(T - T_d) < 3$ °C	8/8 layer
$(T - T_d)$ 3 to 5 °C	Thick layers
$(T - T_d)$ 6 to 10 °C	Thin layers
$(T - T_d) > 10$ °C	No cloud

5.2.2 Cloud formed in precipitation

(a) In rain:
  - (i) Evaporation of precipitation can cool the air to its wet-bulb temperature. A temperature close to the wet-bulb value is reached after about half an hour of very heavy rain or 1–2 hours of moderate rain and results in ragged stratus *pannus*. A rough guide to the base of the stratus *pannus* is:
    2 hours continuous rain — base 800 feet (245 m).
    3 hours continuous rain — base 400 feet (120 m) (no account taken of upslope or advective effects).
  - (ii) Prolonged drizzle and light winds frequently lead to the formation of very low stratus even when upslope motion is apparently negligible.

Advection of warm air across lying/thawing snow often results in stratus forming at or very near the surface.

(b) In snow:
   (i) Melting of falling snow into warm sub-cloud layer will produce an isothermal (along 0 °C isotherm) 600 feet deep after 1 hour, increasing to a maximum of 1200 feet after 4 hours.
   (ii) Extension of the 0 °C isotherm results in a DALR between cooled and unmodified air; at saturation instability causes fractocumulus of a few hundred feet thickness to form near the 0 °C isotherm below cloud base. The modified temperature profile is illustrated in **Fig. 5.1**; the surface temperature will fall as a result of the cooling.
   (iii) If snow reaches the ground, *pannus* forms at or very near the ground.

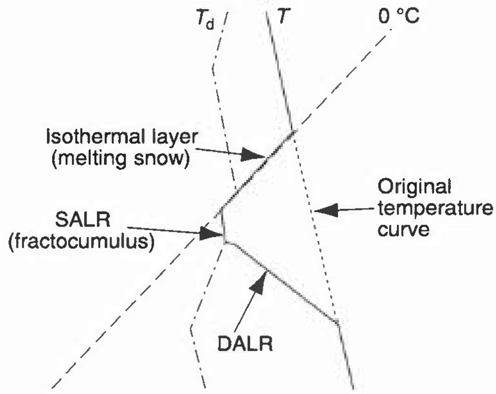

Figure 5.1. Schematic temperature profile produced by melting snow.

5.2.3 Non-frontal medium and high cloud

Most medium and high cloud is associated with mass ascent but may also occur in regions where instability is limited by a capping stable layer. Non-frontal cloud is hard to predict, especially when occurring in thin layers; it is often associated with medium- or upper-level instability or wind shear.

The following may be of help in forecasting its occurrence:

(a) Use of satellite imagery, especially IR (see 10.3.4 and 10.3.5).
(b) Use actual or forecast ascents: $(T - T_d) < 10$ °C, with:
   (i) wind increasing strongly with height;
   (ii) conditional or potential instability at medium/high levels.
(c) Use model cloud or RH fields.
(d) Favourable areas for cloud are:
   (i) cold pools or sharp thickness troughs;
   (ii) on warm side of jet;
   (iii) on, or just to rear of, a sharp upper ridge.
(e) Relate cirrus tops to tropopause (5.2.4.1):
   (i) low tropopause (<30,000 ft, 9 km) — Ci extends to tropopause;
   (ii) high tropopause (>35,000 ft, 10.7 km) — Ci tops a few thousand feet below tropopause.

HWF (1975), Chapter 19.5

5.2.4 Cirrus forecasting

James' technique requires scores to be allotted to a series of questions:

(a) 6–12 hours ahead:
   (i) Is forecast in area of, or just to rear of, an upper ridge?
   (ii) Is forecast area on the anticyclonic side of a 200 hPa jet stream and within 500 km of the jet axis?
   (iii) Is the forecast area within 500 km ahead of a surface warm front or occlusion?

   Score: 0, 0.5 or 1 for 'no', 'uncertain', or 'yes'.
   On adding scores, cirrus amounts are roughly as follows (**Table 5.2**):

Table 5.2.

Total score	0	0.5	1	1.5	2	2.5	3
Amount (oktas)	0	1	2	3	4	5	6–7

(b) 24–36 hours ahead:
   (i) Is the air likely to be moist? — Is the dew-point depression ≤10 °C at or above 500 hPa?
   (ii) Will area be up to 300 miles (480 km) ahead of a surface warm front or occlusion?
   (iii) Will area be on anticyclonic side of a 300 hPa jet and within 300 miles (480 km) of the jet axis?
   (iv) Will area be in, or just to the rear of, a 300 hPa ridge?
   (v) Will area be in a thermal ridge as suggested by the 1000–500 hPa thickness chart?
   (vi) Will the 300 hPa wind over the area be veered from that at 500 hPa by 20° or more?

If two or more affirmative answers are made, cirrus should be forecast 4/8 to 8/8; if not, forecast NIL to 3/8.

James (1957)

5.2.4.1 Tops and bases of cirrus

(i) 50% of all cirrus is likely to be within 5000 ft (1500 m) of the tropopause.
(ii) Thickness is likely to fall within the three ranges: 0–2000 ft, 2000–4000 ft or 4000–6000 ft (0–600, 600–1200 or 1200–1800 m) on roughly an equal number of occasions.
(iii) Mean tops reach the tropopause for tropopause heights up to about 30,000 ft (9 km), becoming relatively lower, until at 39,000 ft (12 km) they are some 4000 ft (1200 m) below the tropopause.
(iv) Mean bases tend to be about 1200 m below the tropopause up to 9 km; at higher levels mean base seems to be in the region of 9 km.
(v) Occurrence of cirrus is favoured in a layer through which wind speed increases rapidly with height **(Fig. 5.2)**.

HWF (1975), Chapter 19.6.4.2

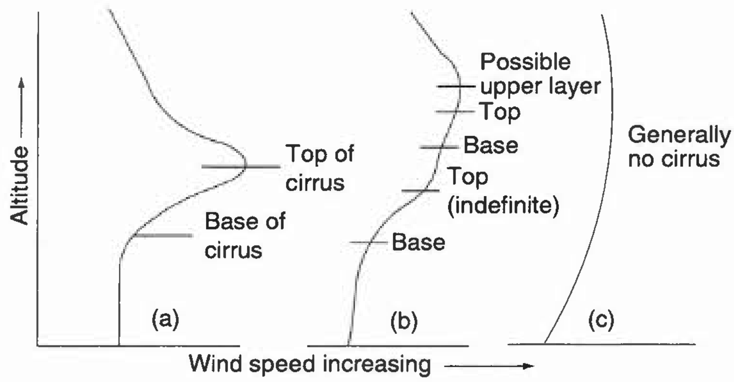

Figure 5.2. Association between cirrus levels and profiles of wind speed.

5.3 Condensation trails

(i) The condensation of water vapour from the exhaust of any aircraft engine can produce long and persistent trails (contrails), the formation and persistence depending on factors such as the ambient air temperature, RH and the amount of exhaust moisture from combusting fuel (see 10.3.5.1).
(ii) The MINTRA line printed on tephigrams gives the critical temperatures for old piston-engined aircraft, based on condensation with respect to ice.
(iii) The continuing usefulness of this line lies in the fact that it represents the limiting temperature above which contrails are unlikely to be formed by any aircraft. In practice, contrails do not normally appear until the ambient temperature is several degrees lower than the printed MINTRA value.
(iv) Recent work by USAF and UK workers has resulted in updated nomograms for the modern turbofan engine, which has a significantly different moisture and heat output.

HAM (1994)

5.3.1 Forecasting contrails

(a) *Using corrections to the MINTRA value*
On a tephigram, draw two lines that are, respectively, 11° and 14 °C colder than the MINTRA critical temperature (T_C) at any level. Plot a representative upper-air sounding. If the air temperature at any level is T, then:

Table 5.3.

		Forecast
If	$T > (T_C - 11)$	contrails unlikely
	$(T_C - 11) > T > (T_C - 14)$	short, non-persistent trails likely
	$T < (T_C - 14)$	long, persistent trails likely

Persistent trails are likely with high humidity and are common when aircraft fly near existing layers of cirriform cloud. Updated corrections show that contrails may form at temperatures warmer than $T_C - 11$ (**Fig. 5.3**); for a dry environment, contrails may not form until significantly cooler. Thus at 200 hPa, with an environmental dew-point depression of 2 °C and a modern, high-bypass engine, trails may be expected at temperatures as warm as $T_C - 9$. With a dew-point depression of 10 °C and a low-bypass engine, trails are not expected until the temperature is less than $T_C - 13$. (Fighters and air-to-air transport refuelling aircraft are examples of the respective engine categories.) If conditions are suitable for contrail formation, then the contrails would be expected to persist for a significant time.

(b) *Graphical method*
Fig. 5.3(a) (after Appleman), includes the effect environmental dew-point depression. Five lines are drawn to show the critical temperature for jet aircraft at different values of dew-point depression. Contrails should always form if the temperature lies to the left of relevant dew-point depression line. No trails are to be expected if the temperature lies to the right of the 0 °C dew-point depression line.

5.3.2 Revised rules for modern engines
Contrail forecasting rules have been revised for two turbofan engine categories (low- and high-bypass) and their associated moisture output and heating; nomograms are illustrated in **Figs 5.3(b) and (c).**

Appleman (1953)
Ferris (1996)
HAM (1994)

5.4 Orographic uplift

5.4.1 Upslope stratus
Factors favourable for development are:
(i) stable atmosphere;
(ii) air near to saturation at lowest levels;
(iii) sloping terrain for the given wind direction;
(iv) moderate wind.

Note that where turbulent mixing maintains extensive low stratus over low-lying terrain, base will be at the mixing condensation level (MCL) **(Fig. 5.4)**; the base on windward slopes is often lower (lifting condensation level, LCL, **Section 4.2**). Also the Föhn effect in precipitation **(Fig. 5.5)** will give a higher base or smaller cloud amount on the lee side (1.3.3.6).

To determine stratus base:
(i) From selected representative sounding modify temperature and moisture profiles near ground to fit local values.
(ii) Find LCL for a series of levels near the surface **(Fig. 5.6)**.
(iii) Upslope base is given by lowest LCL.

(a)

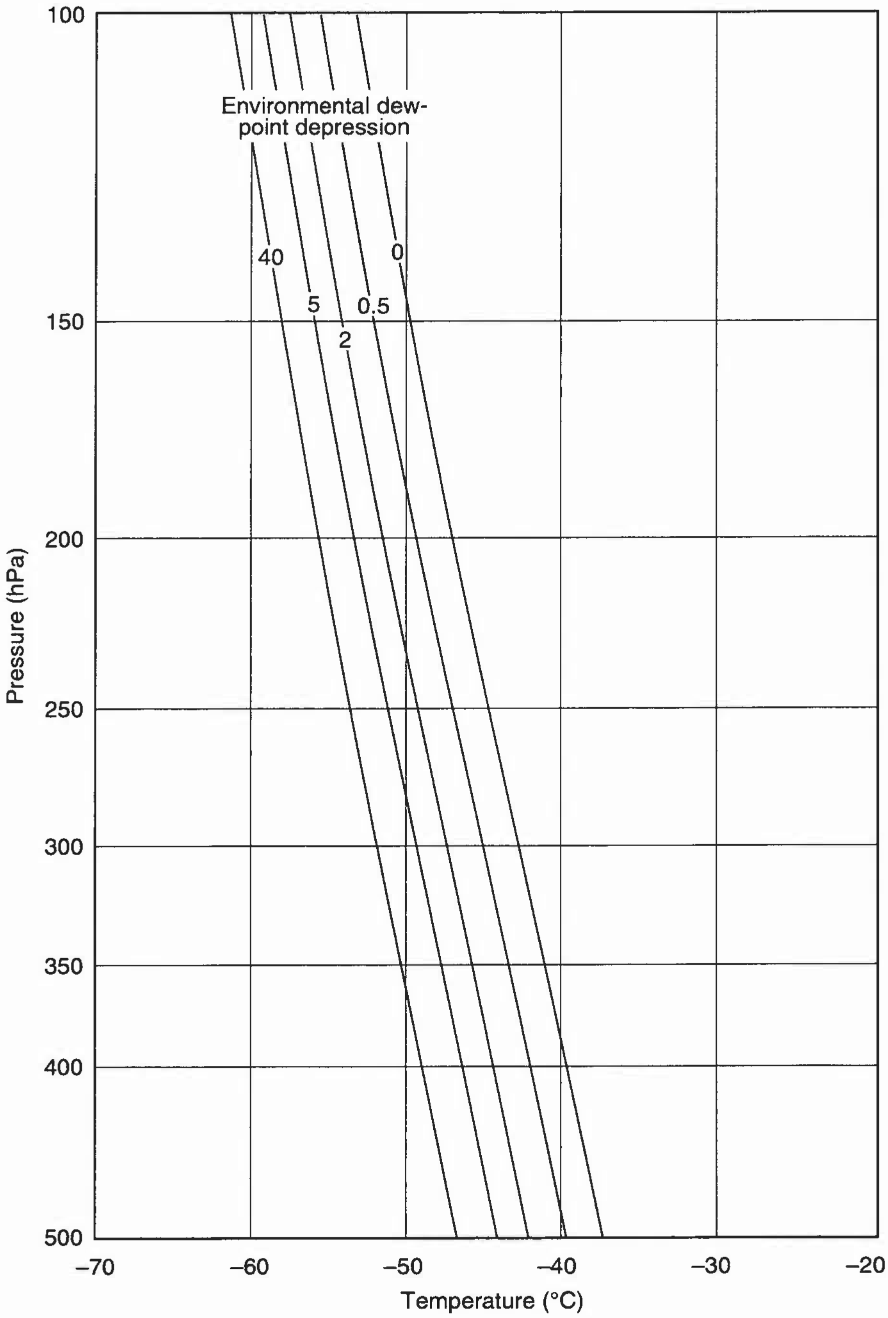

Figure 5.3. (a) The Appleman contrail forecast nomogram.

5.4.2 Orographic clouds

Cap, lenticular, rotor and banner clouds are discussed in 1.3.3 and 10.3.2.

Browning (1975) **Smith (1989)**

5.5 Turbulent mixing

(i) Turbulence tends to establish DALR and constant HMR profiles; RH increases with height, with any cloud base being at the mixing condensation level (MCL) of the layer.

(ii) To assess the MCL requires knowledge of the mixing layer depth, *d*, comparatively easy if top is marked by a sharp inversion. If this is not the case, there are several empirical techniques available, for example:

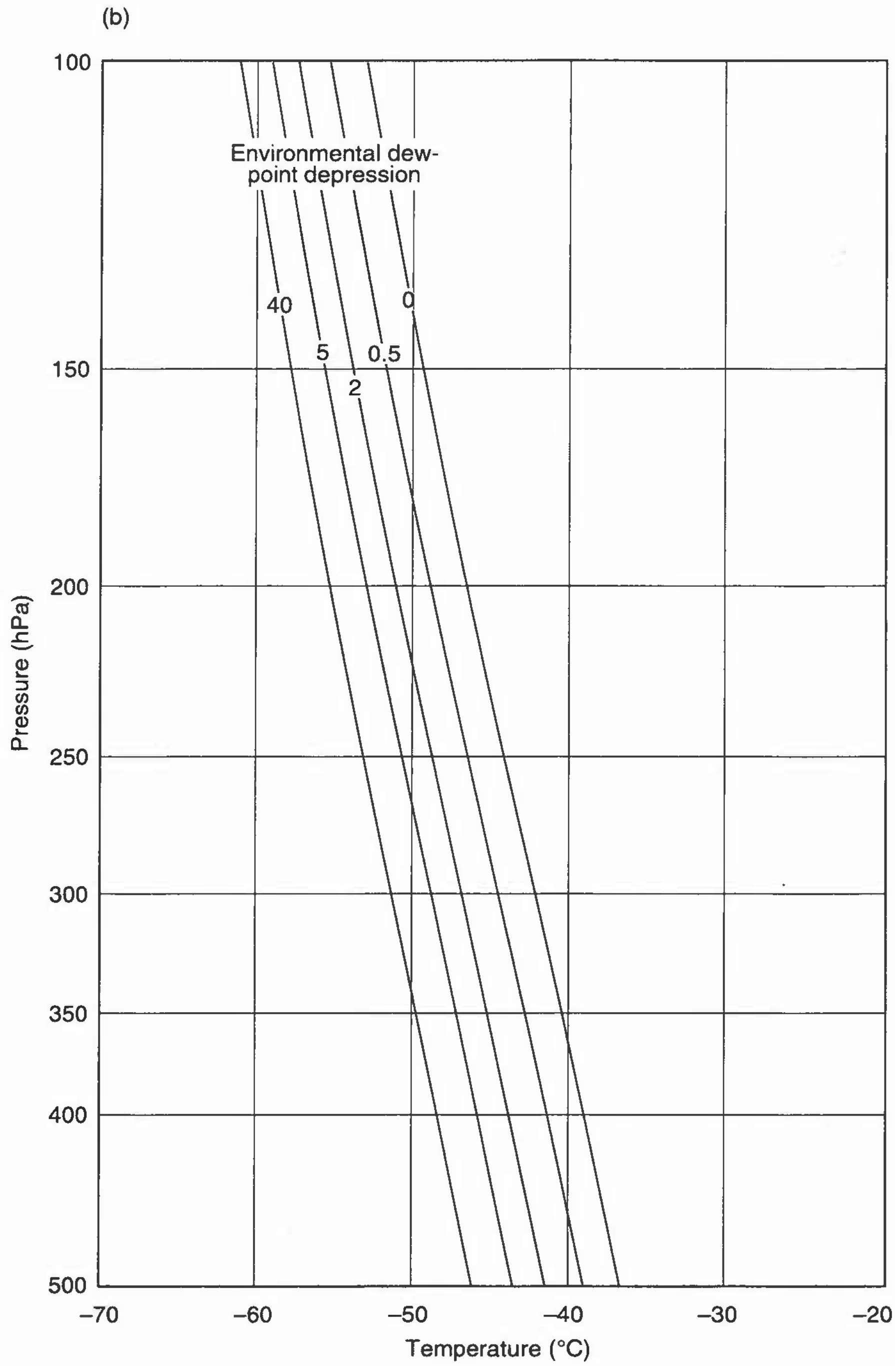

Figure 5.3. (b) low-bypass engine contrail algorithm.

for surface wind, $V_S \leq 16$ kn, $d = 200\ V_S$ feet;
for surface wind, $V_S \geq 16$ kn, d = 3300 to 3600 feet by night
d = 4000 feet by day.

(iii) In view of difficulty in deriving the MCL, locally derived empirical methods may prove more satisfactory.

HWF (1975), Chapter 19

5.5.1 Air-mass stratus

Cooling and turbulent mixing act to lower the boundary layer temperature, with stratus forming below the inversion (**Fig. 5.7**) (see 10.3.3). Three principal regimes identified for the cooling of a turbulent boundary layer below its dew point are illustrated in **Fig. 5.8.** They are:

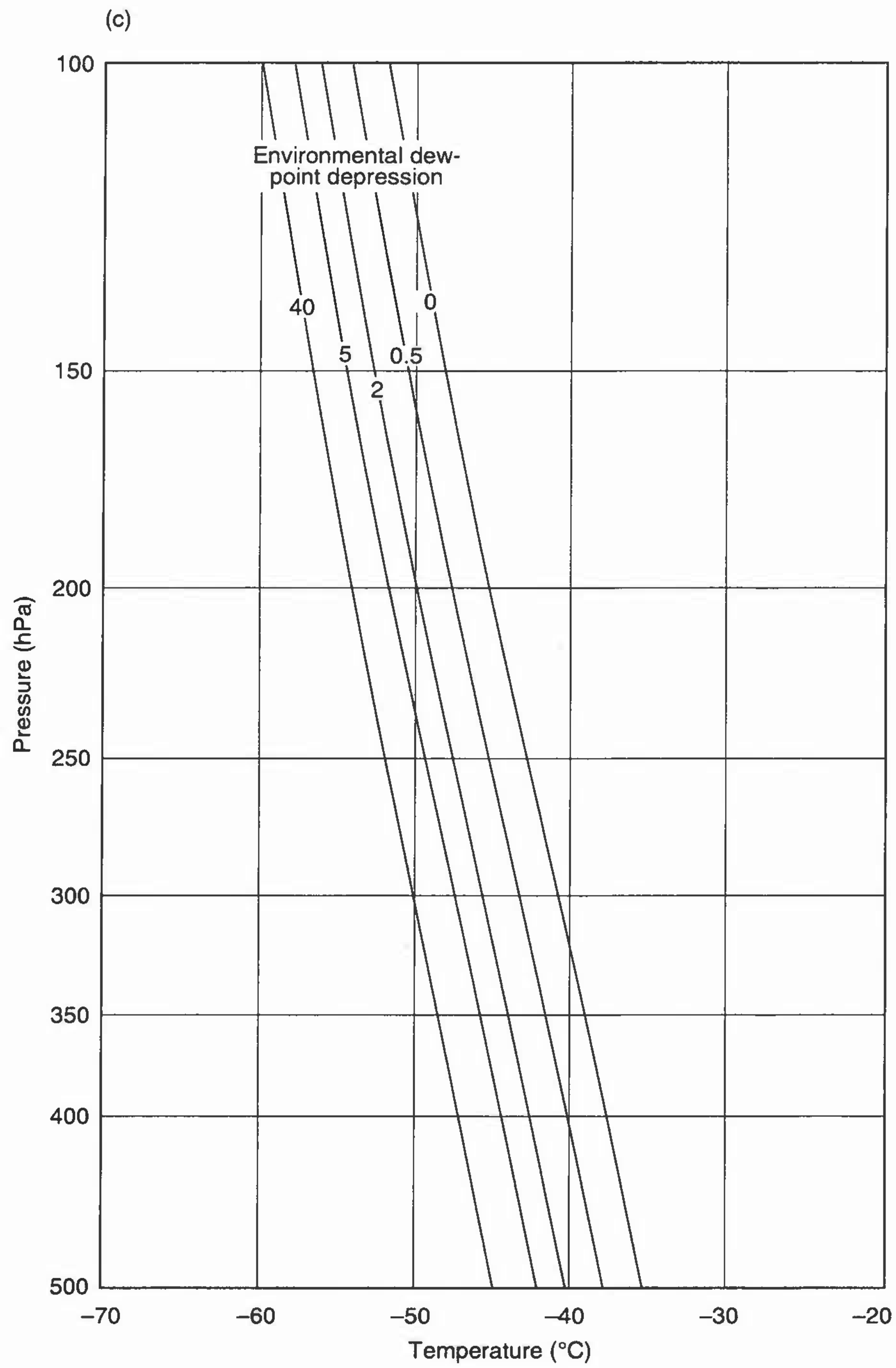

Figure 5.3. (c) high-bypass engine contrail algorithm.

In Fig. 5.8

(a) Moist south-westerlies giving widespread stratus around southern and western coasts, and sometimes sea fog. In autumn and winter inland penetration can give rise to widespread and persistent cover over southern England.

(b) Cold easterlies: in spring and summer relatively warm air advected from the east may be cooled sufficiently by cold sea passage to form stratus on the east coasts of England and Scotland; frequently a problem for airfields in eastern districts.

(c) Winter south-easterly; cloud most likely to form overnight, or when snow is lying. Airfields situated on ground sloping down to the south-east are particularly prone.

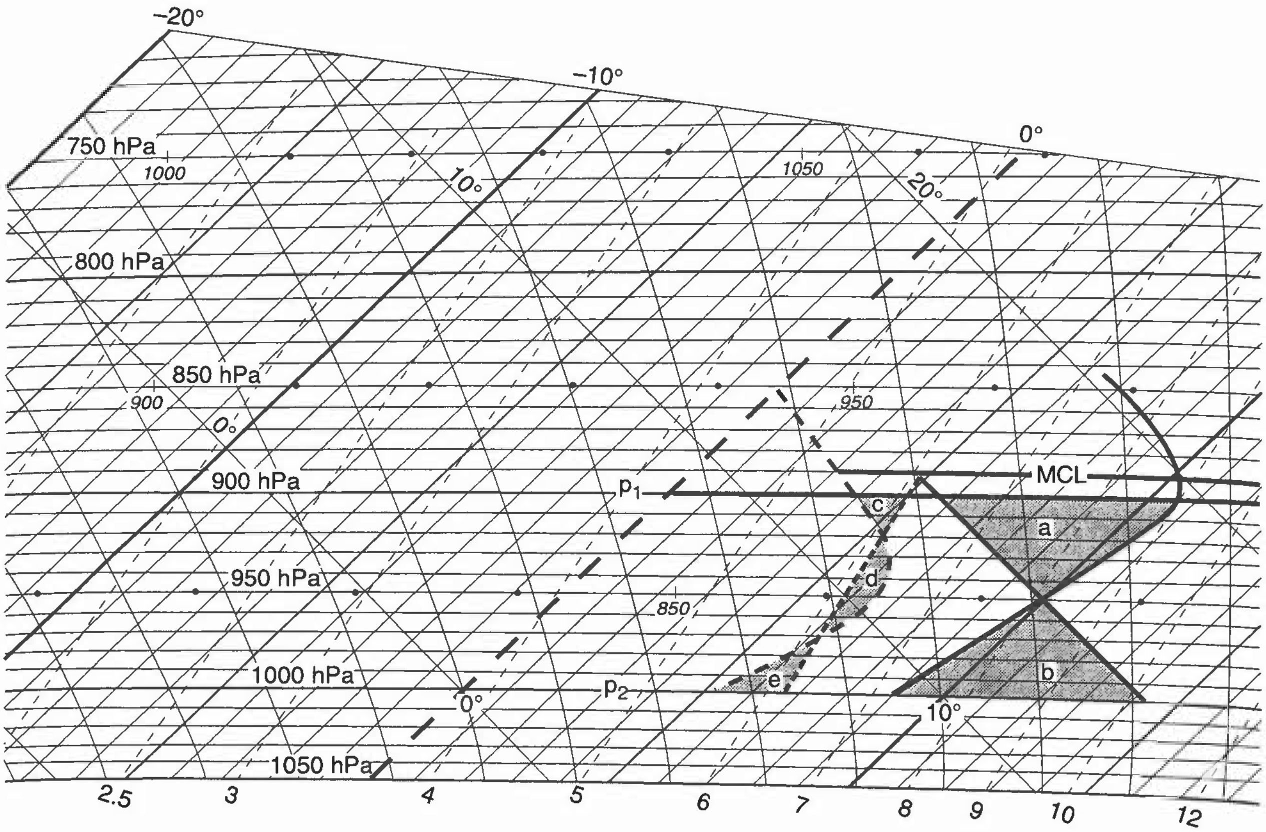

Figure 5.4. Mixing condensation level (MCL) for a layer is found by determining the level at which average HMR and θ_w lines for the layer intersect. The averages are determined by equalizing areas as indicated (area a = b, areas c and e = d).

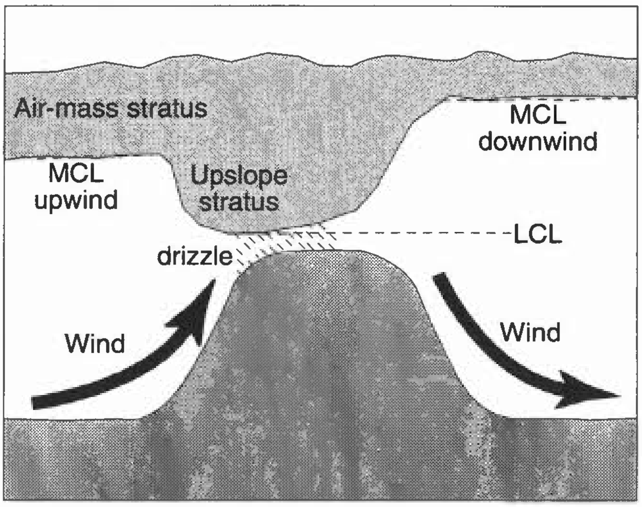

Figure 5.5. The variation of stratus base in response to upslope motion and the Föhn effect.

5.6 Stratus forecasting techniques

5.6.1 Formation of low stratus

To predict the formation of stratus, the following data are needed:

(i) vertical temperature and dew-point profiles of the air mass expected;
(ii) sea temperature (if flow is from over the sea);
(iii) predicted wind field;
(iv) predicted amount and depth of nocturnal cooling overland;
(v) knowledge of the topography.

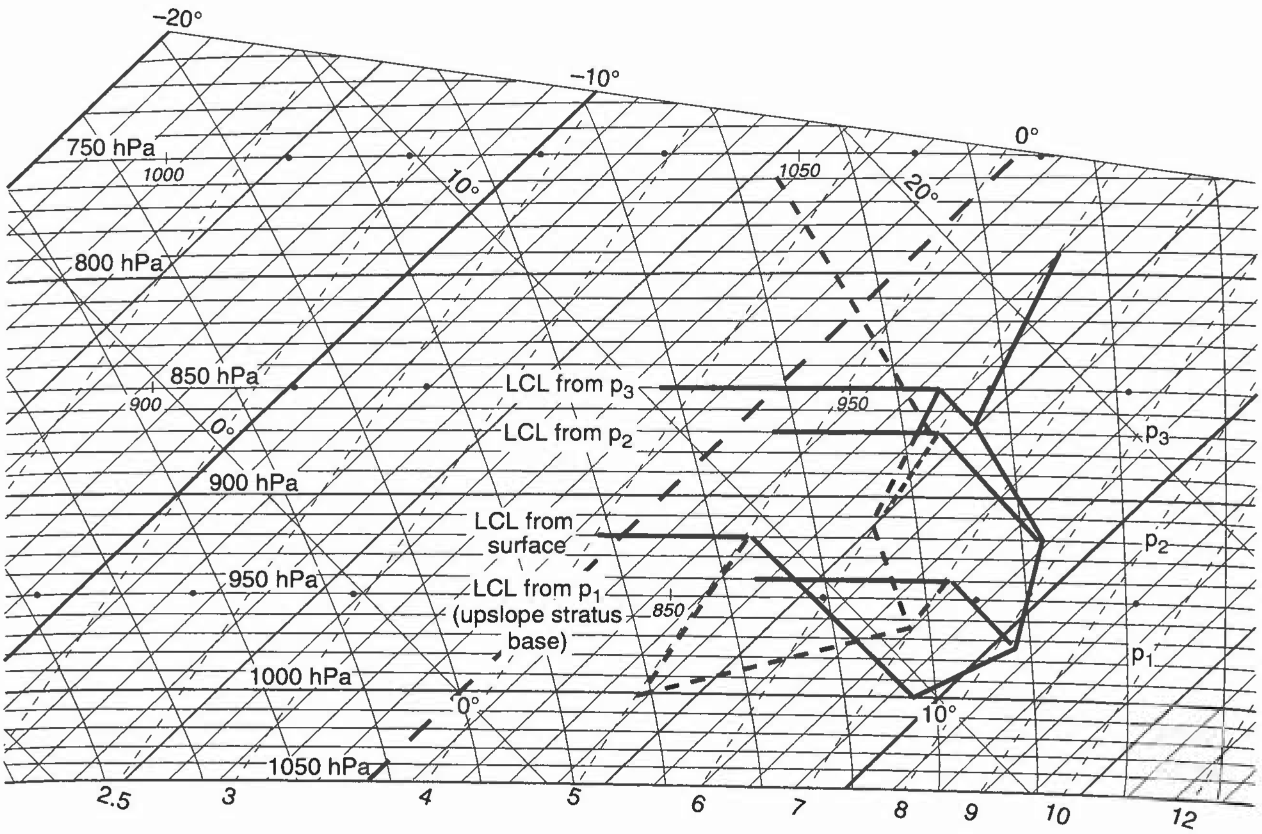

Figure 5.6. Determining the base of upslope stratus (given by the lowest lifting condensation level (LCL) in the lifted layer).

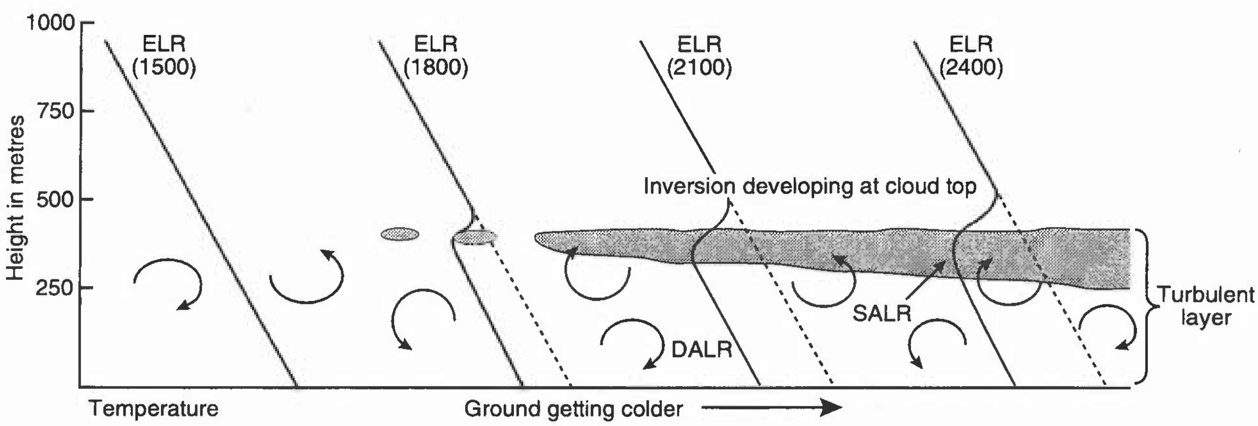

Figure 5.7. The formation of stratus by nocturnal cooling. Temperature profiles are shown every three hours as surface cooling and turbulent mixing act to lower the temperature layer, with stratus forming beneath the inversion.

5.6.2 Forecasting the temperature of stratus formation over land

Formation temperature may be found from a number of sources, the choice of which depends on location and synoptic situation. Stratus may form at:

(i) temperature at which stratus is already present on the chart;
(ii) the stratus clearance temperature earlier in the day (minus 1 or 2 °C);
(iii) the sea temperature if advection is likely;
(iv) the temperature derived from a representative ascent; a technique is given below.

To forecast the temperature at which stratus will form as a result of nocturnal cooling over land **(Fig. 5.9)**:

(i) Obtain the fog-point T_f by Saunders' method (3.3.3.1) from a representative ascent.
(ii) Assess the pressure level at which stratus is expected to form.

(iii) On a tephigram draw the constant mixing ratio line through the fog-point temperature (plotted on the surface isobar).
(iv) From the intersection of this line with the isobar corresponding to the top of the surface turbulent layer (point A), draw a dry adiabatic. The intersection of this adiabatic with the surface isobar indicates the temperature at which stratus will form. This is T_{st}.

The following technique is based on constructing a 'Critical θ_w' (CTW) **(Fig. 5.10)**:
(i) Construct the night cooling curve and find best upwind ascent.
(ii) Draw saturated adiabatic through wet-bulb temperature from the base of the boundary layer inversion to the surface isobar — this is the CTW.
(iii) Mark fog point, F, on surface isobar using Saunders' method.
(iv) Assess mean depth of nocturnal mixing layer by taking depth of 10 hPa for every 6 kn of gradient wind.
(v) Draw top of this mixing layer L′L′. Note where this line cuts environment curve at L; join this point to fog point F.
(vi) This line LF represents the top of any stratus that may form; Normand's point represents the cloud base. Stratus forms when the Normand point passes to a level below this line.
(vii) Monitor cooling by plotting successive Normand points; note where the Normand point seems likely to pass below the line: read off corresponding surface temperature and the time from the cooling curve.

HWF (1975), Chapter 19.6
Warne (1993)

5.6.3 Forecasting the stratus base

The base height is principally a function of wind, which affects the efficiency of turbulent mixing, and RH, which determines the height of the MCL. Generally, the stronger the wind, the higher the base, as given by the guide:

h (ft) = 75 V_S, for forecast surface wind in knots.

The surface temperature/dew-point difference may be used: $h = \alpha(T - T_d)$, α = 420 feet (130 m), although work at Shawbury has given values between 395 and 450 (the latter for tropical maritime air). However, the surface dew-point depression may not be representative of the mixing layer, and base estimates will be under- or over-estimated depending on whether the ground is cooling or warming. The observed base of stratus that has already formed upwind may be a more useful guide, adjusted to local effects.

HWF (1975), Chapter 19.6

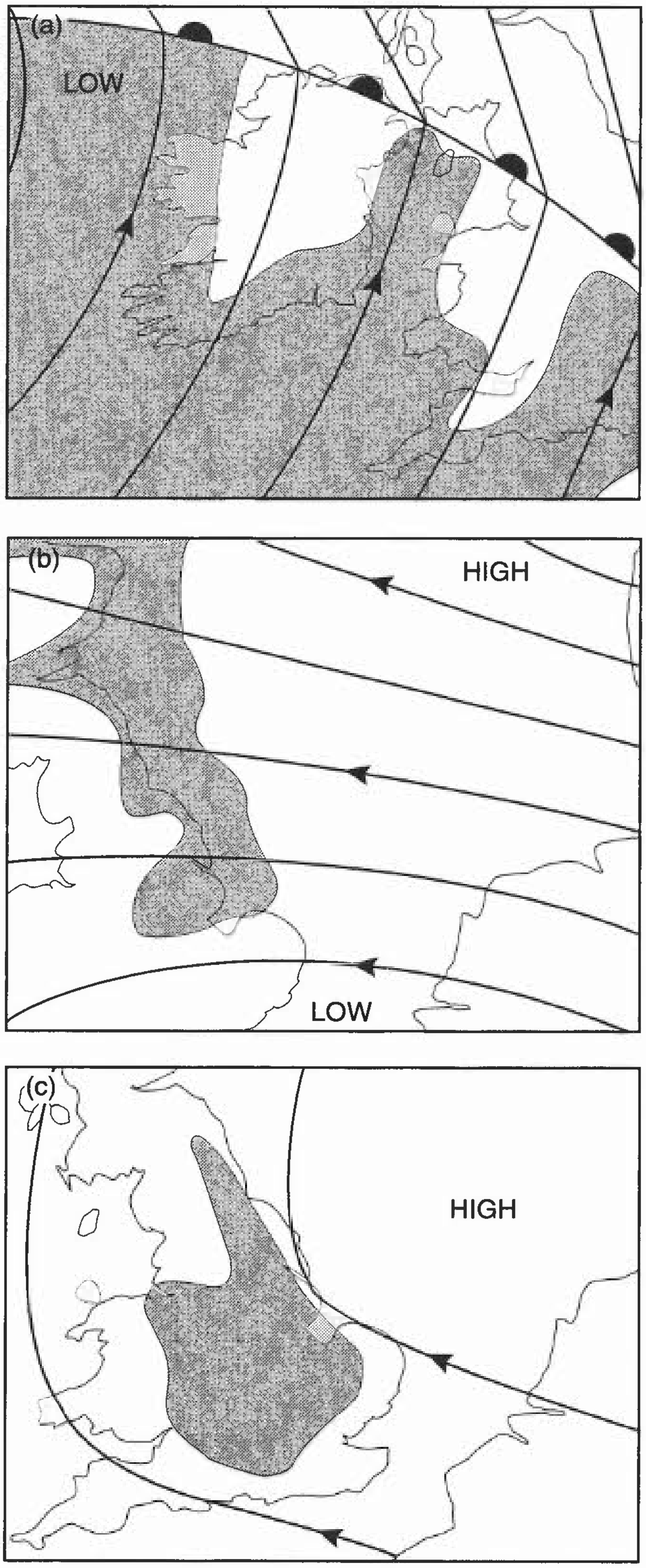

Figure 5.8. Areas of the British Isles prone to stratus (stippled) in the three principal regimes, (a) moist south-westerly, (b) cold easterly, and (c) winter south-easterly.

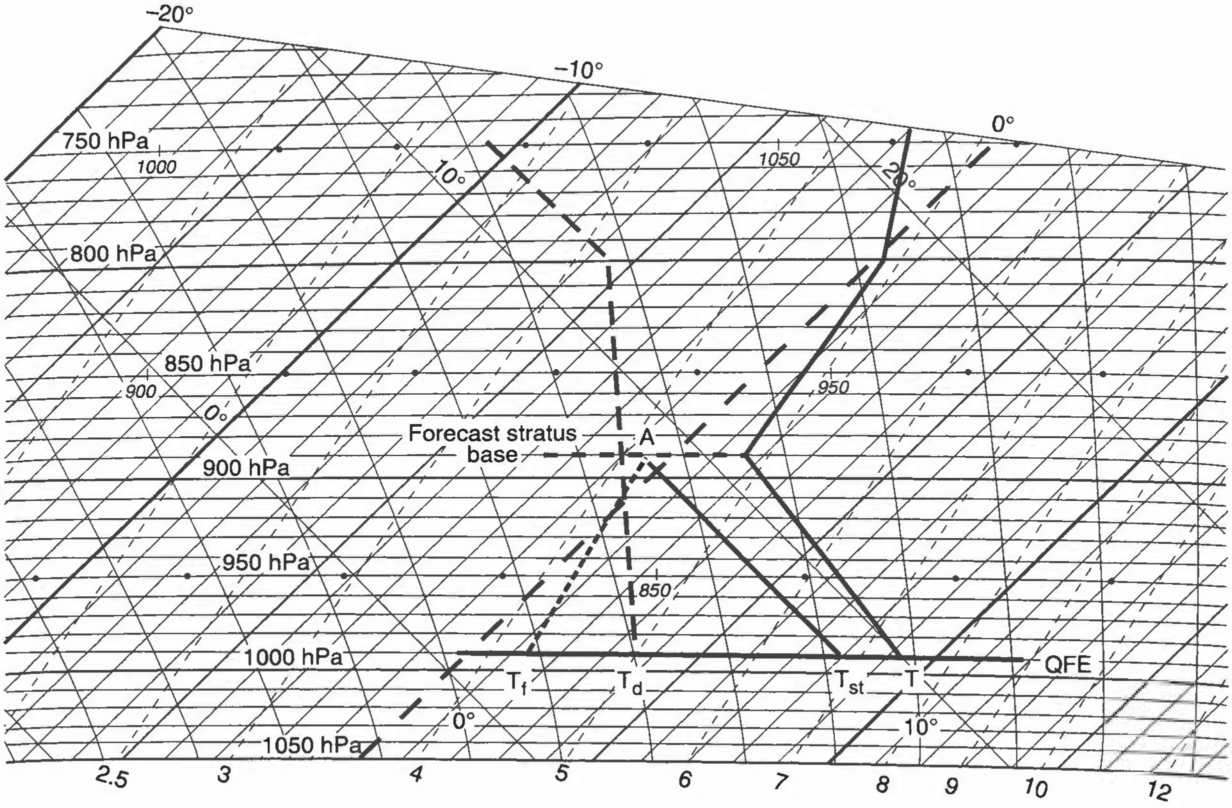

Figure 5.9. Determination of the stratus formation temperature (T_{st}) from a representative ascent. See text for details.

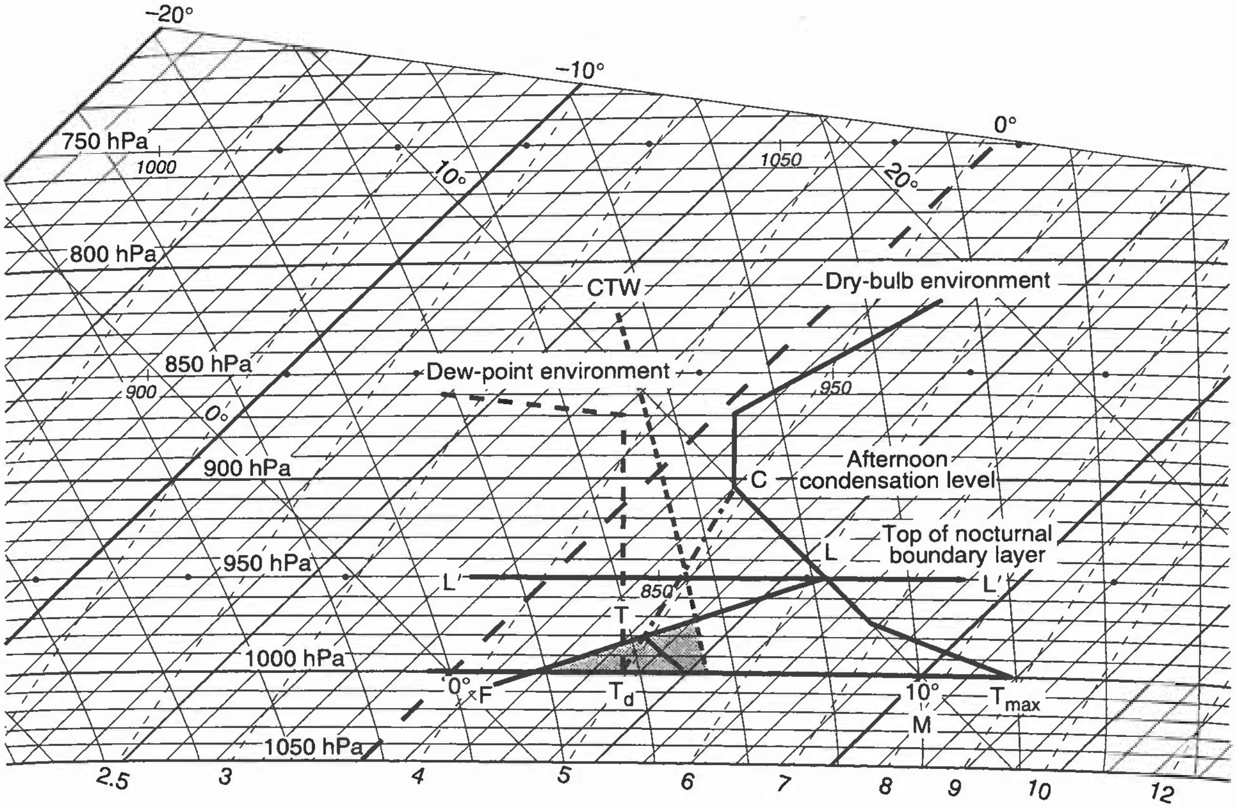

Figure 5.10. The stratus construction lines (on a tephigram) on an ideal sounding with a gradient wind of 30 kn. True stratus forms in the shaded area, 'cumulostratus' in the area between ML at the CTW.

5.6.4 Forecasting stratus tops

The top of the mixing layer will mark the stratus tops; it should be noted that, in most situations, stratus is likely to be no more than a few hundred feet thick. The 'critical θ_w' technique also gives stratus tops.

5.6.5 Forecasting the advection of stratus from the sea

(i) If there has been little change in the synoptic situation during the day, the temperature at which stratus cleared in the morning (minus 1 or 2 °C) is a good estimate of the threshold temperature at the coast when the stratus will start again to move inland from the coast. (If coastal temperature is not known use the (representative) sea temperature.)

(ii) The movement of stratus inland can be forecast from the direction and speed of the 10 m wind at the time when convection and turbulence are still operative in the lowest layers, and before the surface temperature begins to fall towards evening. Valleys aligned at right angles to the coast facilitate deep inland advances.

(iii) If there is upslope motion, stratus may form over higher ground before the main cloud sheet spreads in from the coast (giving the impression that the movement inland occurs at greater than the geostrophic speed).

(iv) The level at which stratus will form can be determined from a representative vertical profile of temperature and dew-point. Use a Normand construction to determine the condensation levels of air lifted from several levels. The lowest condensation level represents the pressure at the base of any orographic stratus.

(v) Studies around the Eden estuary (east Scotland) have shown that 'haar' may recede with the ebbing tide as estuarine muds and sands are exposed and warm rapidly; the haar advances with the next tide.

Alexander (1964) **Lamb (1945)**
Sparks (1962)

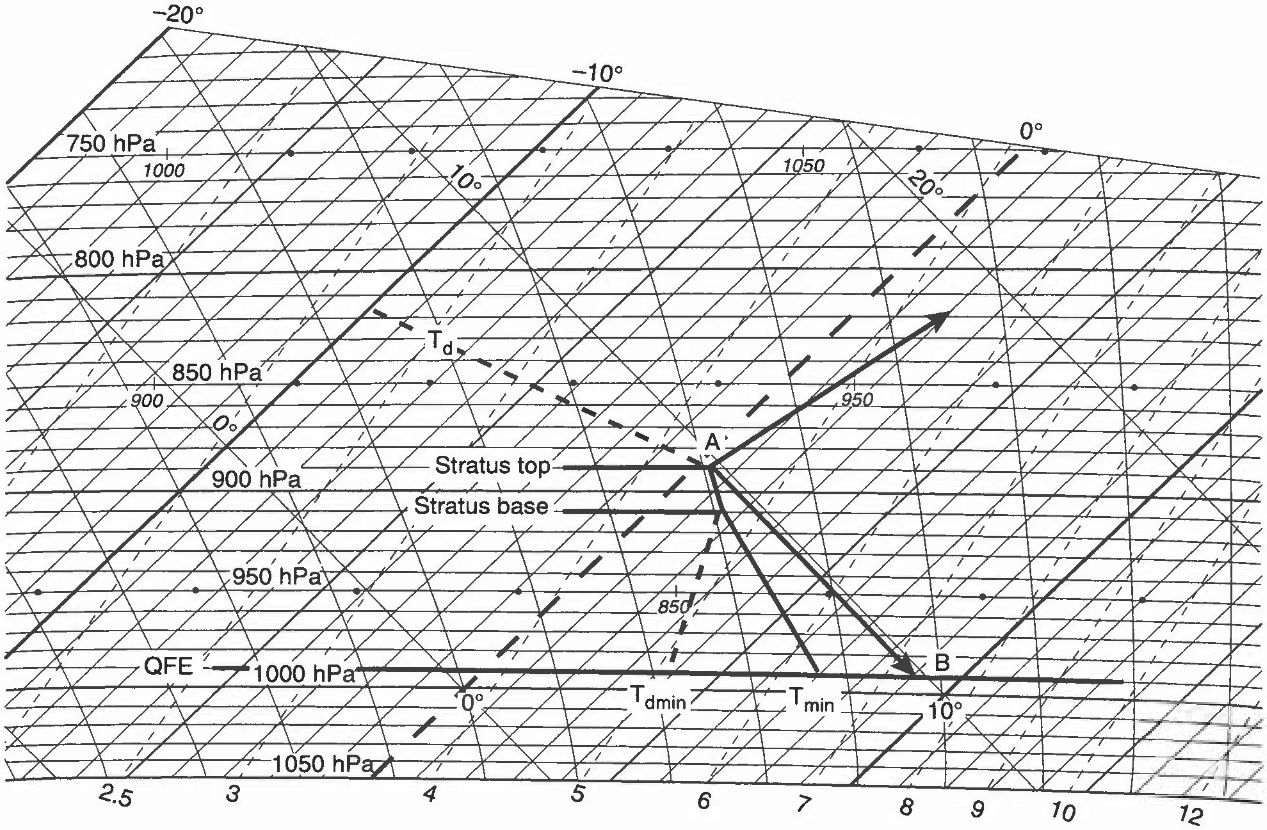

Figure 5.11. (a) Determination of the stratus clearance temperature from a representative ascent: construct a dry adiabat from the level of the stratus top on the temperature curve (point A) to the surface pressure level QFE (point B); the temperature at B is the stratus clearance temperature.

5.6.6 Stratus clearance

The three main clearance mechanisms are: *Insolation; Increased wind; Advection of drier air*

(a) *Insolation* — from a representative ascent construct a dry adiabat from the level of the stratus top to surface; this is clearance temperature. If a day heating curve is constructed assuming cloudy conditions, the time of clearance will be that at which stratus clearance temperature is reached (point B in **Fig. 5.11**). In winter stratus often persists most of the day, clearing briefly at the time of maximum temperature.

(b) *An increased wind speed* — will deepen the mixing layer, mixing in drier air from above.

(c) *Advection of drier air* — in such cases the rear edge may be followed on satellite imagery and/or charts. One common example is the clearance of North Sea stratus by the backing or veering of the wind to give a shorter sea track from the continent.

Mansfield (1988)

5.7 Stratocumulus: physical and dynamical processes of formation and dissipation

Through its significant effect on the radiation balance Sc affects boundary layer structure and hence fog formation and dispersal, maximum and minimum temperatures, etc. Since Sc still presents a formidable forecasting challenge, it is important that forecasters be aware of the latest thinking on the physical and dynamical processes of formation and dissipation so as to be able to base Sc forecasts on as much physical insight as is currently available (see 10.3.1.3).

5.7.1 Features of formation

Sc formation is usually associated with either:

(i) cooling/moistening of the boundary layer or

(ii) spreading of Cu beneath an inversion.

Figs 5.12 and 5.13 show five regions of interest:

A— just above cloud; air is warm and dry through subsidence, thus stable with little turbulence.

B— at cloud top; marked inversion and hydrolapse. Very stable; local perturbations strongly damped although small-amplitude gravity waves are often present. Wind shear, sometimes large, is confined to this inversion layer.

C— within cloud; well mixed, constant WBPT. Liquid-water content slightly below the adiabatic value due to entrainment. Aircraft icing a danger with temperatures in the range 0 to −15 °C.

D— cloud base; transition to clear air poorly defined, no strong temperature gradients. Perturbations can grow as is evident from rolls frequently observed.

E— below cloud base; well mixed with same WBPT as in cloudy layer. Potential temperature profiles show discontinuity but within- and below-cloud regions may be treated as one.

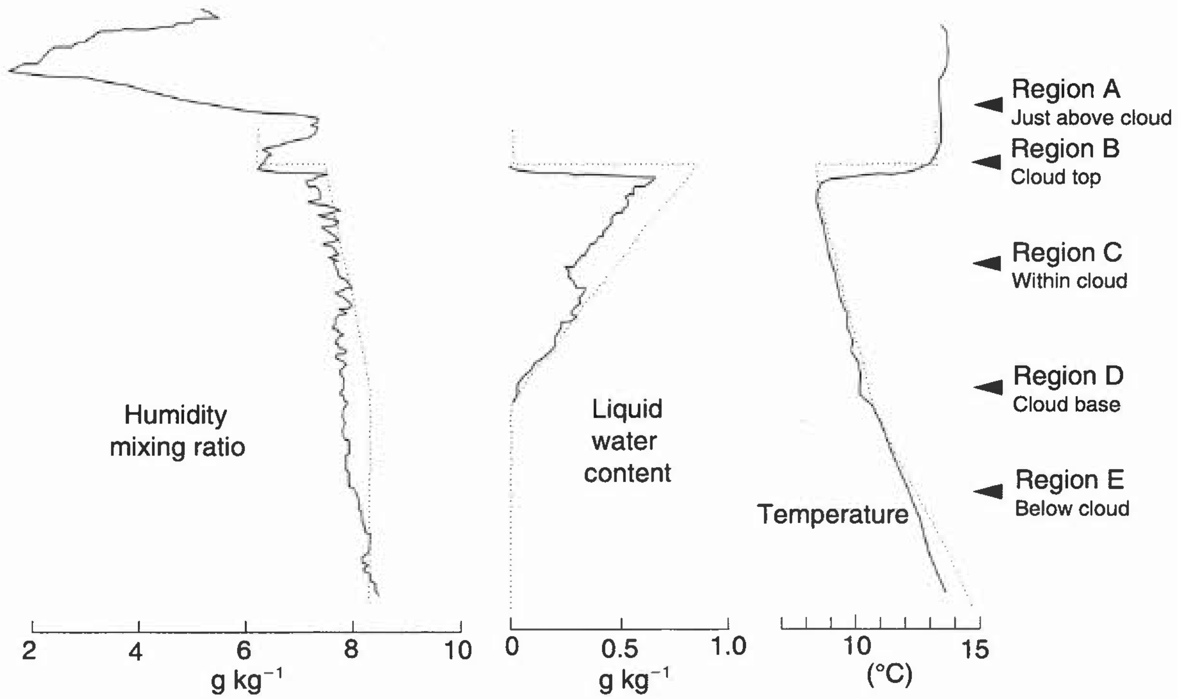

Figure 5.12. An example of the humidity mixing ratio, liquid water content and temperature profiles in stratocumulus. The solid lines show the structure as measured by the Meteorological Research Flight C-130 aircraft. The dotted lines show an idealized representation for stratocumulus.

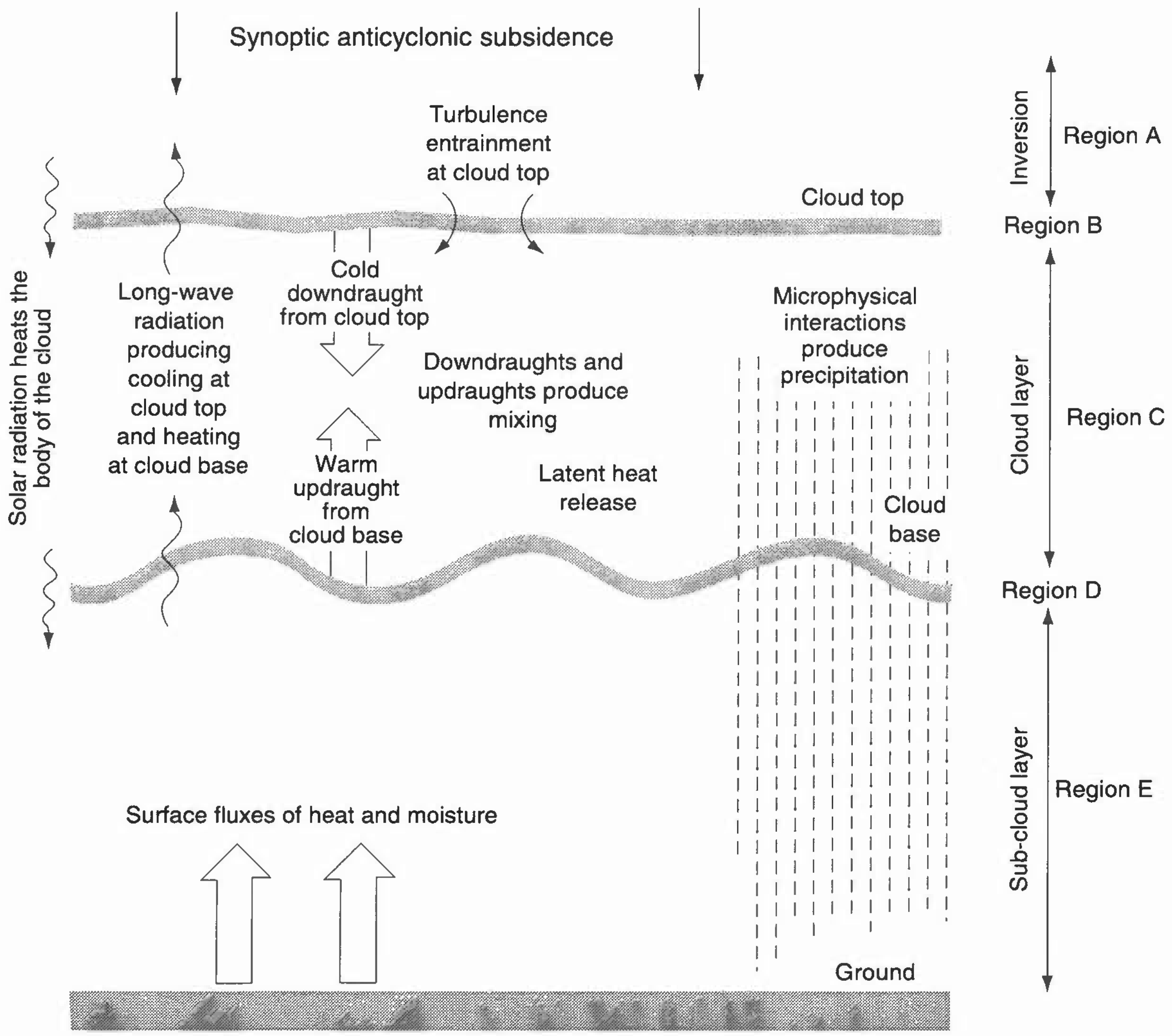

Figure 5.13. Summary of physical processes important to the development of stratocumulus.

5.7.2 Control mechanisms of Sc development

Processes include:

(a) synoptic-scale subsidence (tending to lower cloud top)
(b) radiative effects important (lifetime of Sc several days)
(c) turbulent mixing raises cloud top
(d) detailed structure defined by microphysical processes.

In more detail:

(a) Radiative heating and cooling generates convective instability, mixing and thickening clouds; this is opposed by large-scale subsidence. Thus cloud layer will basically rise and fall depending on which mechanism temporarily dominates.
(b) (i) When cloud thickness is >400 ft (150 m), it becomes important to consider the long-wave radiation balance between cloud top, base and within the cloud layer. Radiative cooling rate at cloud top is about 5 °C h^{-1}, with a slight warming at the base since surface is warmer than cloud base.
(ii) Of the solar radiation penetrating, 13–15% is absorbed; remainder reaches ground. In June in the UK balance of solar radiation emitted/absorbed nearly equal. Sc will display significant diurnal changes.
(c) Turbulence and entrainment generate convective instability; long-wave cooling and solar heating will rarely be in exact balance and cloud experiences a net cooling.
(d) Aircraft measurements show there to be little variation of droplet concentration with height, although drops become bigger towards cloud top. Whether there is drizzle at the ground will depend on the humidity of air below cloud base and local orographic effects.

Fig. 5.14 is a summary of the physical processes important to Sc development.

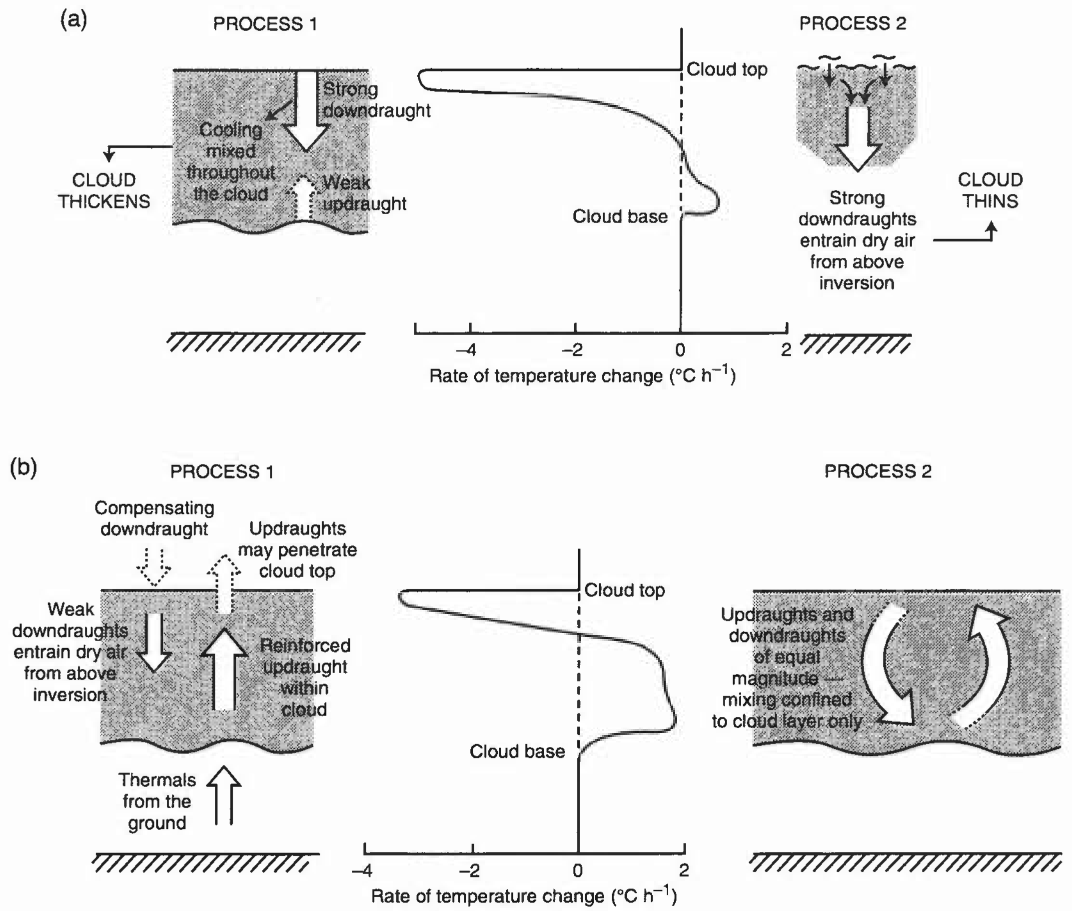

Figure 5.14. (a) Outline sketch showing the effect of the two competing processes on a sheet of nocturnal stratocumulus: Process 1 — radiative cooling at cloud top produces downdraughts which mix cold air throughout the cloud layer and cause stratocumulus to thicken, and Process 2 — strong downdraughts entrain air which when mixed with cloudy air may disperse the stratocumulus. (b) Outline sketch showing the effect of the two competing processes on a sheet of daytime stratocumulus: Process 1 — updraughts are enhanced due to warming and may penetrate the cloud top causing compensating downdraughts to entrain dry air, and Process 2 — updraughts and downdraughts are of equal magnitude confining mixing to the cloud layer only.

5.7.2.1 Night-time effects (Fig. 5.14(a))

Two competing processes are to be considered:

Process 1

(i) At night there is a net cooling in cloud layer.
(ii) Since water vapour content remains basically constant, cloud formation is enhanced.
(iii) Net effect: cloud becomes denser and cloud base lowers.

Process 2

(i) Net cooling at the top changes stability of layer and leads to enhanced instability, principally through the presence of stronger downdraughts.
(ii) Stronger downdraughts increase entrainment of dry air from above inversion,
(iii) If air above the inversion is sufficiently dry the Sc disperses through mixing.

From forecasting point of view:

Process 1 — precise details of cloud layer are unimportant once persistence is assured.
Process 2 — basis of James' rule (5.8.2.1): if $D_m > D_c$ then air above the cloud layer is dry enough to evaporate the cloudy layer completely; applicable over land only. Not very reliable because of difficulty of making measurements accurate enough to distinguish between the competing processes.

5.7.2.2 Daytime effects (Fig. 5.14(b))
Starting with night-time Sc two processes can be envisaged as the sun warms cloud and ground:

Process 1

(i) Both main body of cloud and the ground warm, achieving a radiation balance.
(ii) The turbulence structure changes; updraughts are enhanced by thermals from the ground; downdraughts weaken.
(iii) If the updraughts are strong enough, they may penetrate inversion, inducing compensating downdraughts and forcing dry air from above the inversion into the cloud layer.

If updraughts do not reach cloud top then no clearance possible; if they do penetrate inversion, air above top may not be either dry enough or warm enough to induce clearance. It is crucial that updraughts should be strong enough and air sufficiently warm and dry in order to clear the cloud. This delicate balance is reflected in Kraus' rule (5.8.2.2) which, however, only uses temperature change across the inversion (although, in subsiding air, hydrolapse and temperature inversion are often closely linked).

Process 2

(i) Insolation heating and long-wave cooling will sometimes be comparable, inducing updraughts and downdraughts of similar strength, thus confining mixing to the cloud layer. Sub-cloud moisture then no longer enters the cloud so cloud base will rise and cloud will thin and possibly disperse. A weak inversion will develop beneath cloud base.
(ii) Daytime observations may then suggest that the main cloud layer is beginning to thin and will eventually clear.

This section has concentrated on radiatively-driven convection; other forms of mixing (e.g. through wind shear) may also be important, but the cloud will then have a different structure and behaviour.

Bennetts et al. (1986)
James (1959)
Kraus (1943)

5.8 Non-frontal stratocumulus

In the formation of stratus, the turbulence is generally purely mechanical; in the case of stratocumulus, convection and radiation often play major roles.

5.8.1 Formation and dispersal
Two main categories of Sc are:

(a) Sc formed by spreading out of Cu — this is a common occurrence in subsiding Pm air masses, particularly on the eastern flank of anticyclones (4.4).
(b) Anticyclonic Sc:
Formation: is by mechanical turbulence in the mixing zone, by convection due to surface heating, or a combination of both. Convection is particularly important when cold air advects over warm sea and the inversion limits convective depth.

Forecasting notes:

(i) Advect the cloud edge with the wind at cloud level.
(ii) Use satellite imagery.
(iii) Best to forecast NIL or 8/8.
(iv) Sheet usually reduces insolation and inhibits night cooling.
(v) May be very persistent, especially in winter.

Dispersal: prediction is difficult, especially in winter.
Forecasting notes:

(i) Look for rear edge and advect with wind at cloud level, or by continuity (from satellite pictures or upwind observations)
(ii) If surface temperatures rise to give DALR profile to cloud top, break-up may start but re-form when temperature drops.

(iii) Continued subsidence may lower the inversion sufficiently to bring down drier air and disperse the cloud (unusual).

(v) James', and Kraus' rules are two forecasting techniques discussed next; neither is particularly reliable.

Bennetts et al. (1986)

5.8.2 Dispersal of stratocumulus

5.8.2.1 Nocturnal dispersal of stratocumulus over land (James' rule)

Only use this technique if the stratocumulus sheet is bounded by a dry inversion, there is no surface front within 400 miles and the cloud sheet is extensive, giving almost complete cloud cover (>6/8 for 2 or more hours).

The cloud will break if: $D_m > D_c$

where:

D_m is the maximum dew-point depression (°C) in the 50 hPa layer above cloud

D_c is the value given in the **Table 5.4** below, in which

b is the difference (g kg^{-1}) between HMR at top and bottom of the 50 hPa layer below cloud

z is the cloud thickness (hPa).

Table 5.4. Values of D_c (°C) for use with James' rule

z	b (g kg^{-1})					
(hPa)	0.25	0.50	0.75	1.00	1.25	1.50
10	—	—	1	3	6	8.5
20	0	2.5	5	8	10	13
30	4	7	9	12	14.5	17
40	9	11	14	16	19	21
50	13	15	18	20.5	23	26
60	17	20	22	25	27	30
70	21	24	26.5	29	32	34

Note: a linear hydrolapse in the layer is assumed.

James (1959)

5.8.2.2 Dispersal of stratocumulus by convection (Kraus' rule)

A cloud layer will not disperse by convective mixing with the air above if the pressure (hPa) at the cloud top is less than P_c, as given below. (If the pressure at cloud top is greater than P_c the cloud may or may not disperse.)

$$P_c = P + a(P_0 - 1000)$$

where P_0 is the surface pressure (hPa) and P and 'a' are given in **Table 5.5** below.

Kraus (1943)

Table 5.5. Values of P and 'a' (for use with Kraus' rule) for given cloud-top temperatures (water or ice cloud) and strength of inversion.

	Temp. at cloud top (°C)	Magnitude of inversion containing the cloud layer (°C) 10		8		6		4		2	
		P	a	P	a	P	a	P	a	P	a
Water cloud	20	833	0.80	861	0.83	891	0.87	924	0.90	960	0.95
	10	803	0.75	834	0.79	869	0.82	906	087	951	0.93
	0	755	0.67	789	0.71	830	0.76	877	0.82	932	0.90
	−10	680	0.56	719	0.60	765	0.66	823	0.73	898	0.84
Ice cloud	0	779	0.71	812	0.75	850	0.79	891	0.85	941	0.91
	−10	702	0.59	739	0.63	786	0.69	839	0.76	908	0.85
	−20	586	0.45	628	0.49	679	0.54	747	0.62	841	0.74
	−30	451	0.30	489	0.34	540	0.38	613	0.45	728	0.58

5.9 Precipitation from layered clouds

5.9.1 Frontal and non-frontal precipitation

5.9.1.1 Definition of intensities of (non-showery) precipitation (UK Met. Office)

Table 5.6. Definition of intensities of (non-showery) precipitation (UK Met. Office)

(a) *Rain*	
Slight	<0.5 mm h^{-1}
Moderate	0.5 to 4.0 mm h^{-1}
Heavy	>4.0 mm h^{-1}
(b) *Snow*	
Slight	<0.5 cm h^{-1}
Moderate	about 0.5 to 4.0 cm h^{-1}
Heavy	>4.0 cm h^{-1}

Note: 1.25 cm of fresh snow ≈1 mm water.
1 foot of fresh snow ≈1 inch of rain.

Observer's Handbook (1982)

5.9.1.2 Precipitation in frontal depressions

Synoptic observations, satellite imagery, continuity, tephigrams and upper-air information will confirm the presence or otherwise of any jet streams or upper troughs that will encourage a front to become more active. The amount and intensity of the precipitation depends on:

(i) large-scale ascent of air;
(ii) the continued availability of moist air, originating at low levels;
(iii) the release of potential instability aloft;
(iv) orographic enhancement over high ground.

Frontal precipitation often occurs in bands, usually but not always parallel to, or coincident with, one of the surface fronts. (Bands of rain ahead of a cold front may penetrate well into the warm sector or even ahead of the warm front. These bands are mesoscale phenomena.) (Chapter 7).

Browning (1985)
Browning et al. (1974) **Browning et al. (1975)**

5.9.1.3 Non-frontal precipitation
If non-frontal precipitation is tied to a significant (organized) feature, a surface trough or polar low, for example, then a reasonable forecast of precipitation changes can be made, often by advecting the precipitation area with the 700 hPa wind.

5.9.1.4 Quantity of precipitation
Synoptic data and radar rainfall displays are the best guide to timing the onset and cessation of precipitation and for assessing development, movement and intensity of precipitation areas.

5.9.2 Forecasting drizzle
(i) Precipitation from stratiform cloud is related more to the microphysics than broader-scale dynamics.
(ii) Drizzle is often associated with haar/sea fog (fret), affecting exposed coasts, being cleared inland by insolation and, near coasts, by advection. It thus tends to be more frequent during the latter part of the night and early morning.
(iii) For drizzle to reach the ground it must fall from stratus cloud with its base less than about 1500 ft (450 m) and a minimum depth of 2000 ft (600 m). The dew-point depression in the air below cloud should be less than 2 °C, otherwise the very small drizzle droplets (0.2–0.5 mm diameter) will evaporate.
(iv) Clouds of thickness <7500 ft (2250 m), cloud-top temperatures <−12 °C and bases colder than 0 °C rarely give rain **(Fig. 5.15)**.
(v) Satellite imagery can help to identify areas of colder cloud tops where precipitation is more likely. Heavy drizzle mostly occurs within clouds that are covering high ground but radar imagery often underestimates drizzle intensity due to the 'overshooting' of the beam (10.6).

HWF (1975), Chapter 19.7

5.9.3 Depth of cloud for precipitation
Fig. 5.15 gives a guide to the intensity of precipitation at ground level, over flat terrain, associated with different thicknesses of cloud; it thus defines criteria that distinguish precipitating from non-precipitating cloud although cloud-top temperature is not considered. It is constructed for layer clouds in which the difference in water content between the base and the top is over 1.5 g kg^{-1}. Precipitation intensity in polar air masses is likely to be enhanced by ice processes even in quite shallow clouds (about 2000 feet).

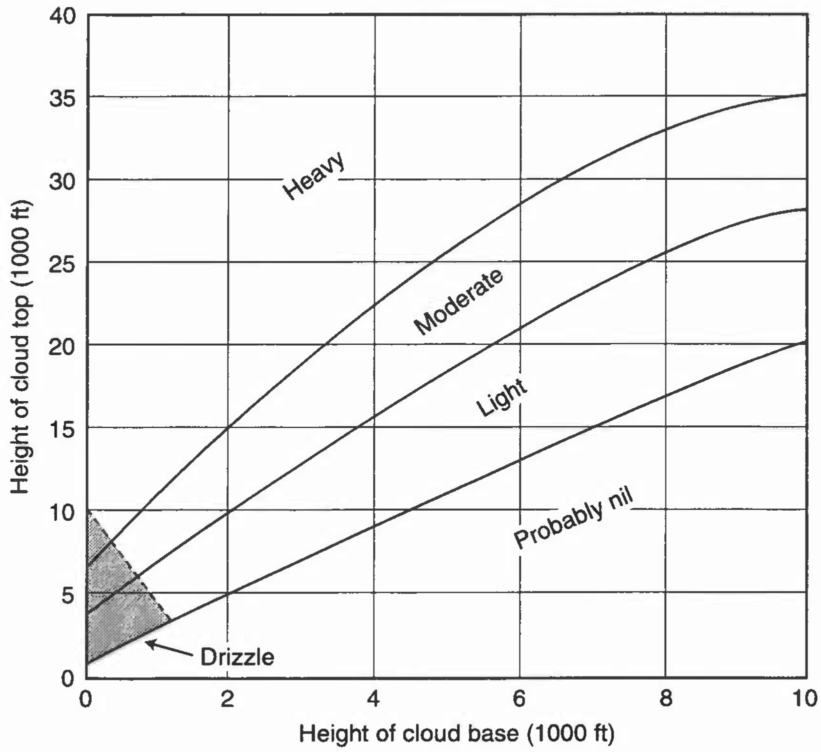

Figure 5.15. The depth of cloud related to intensity of precipitation. The stippled area indicates the conditions for the precipitation to be mainly drizzle.

In summary:

(i) There is a high probability of precipitation from clouds with thickness >7500 feet (2250 m) with cold tops (<−10 °C).

(ii) Deep clouds with bases <3300 feet (1 km) above the 0 °C isotherm — precipitation likely, while such clouds with bases >3300 feet above the 0 °C isotherm — precipitation unlikely.

HWF (1975), Chapter 19.7

Whether there is drizzle at the ground from Sc will depend on the humidity of air below cloud base and local orographic effects (**Table 5.7**).

Table 5.7. An estimate of stratocumulus depth to produce drizzle at cloud base (taking cloud top at −5 °C)

Air mass	Minimum depth to produce drizzle at cloud base (m)	(ft)
Clean maritime	500	1600
Maritime	1000	3300
Continental	2000	6600
Industrial continental	2500	8200

Bennetts et al. (1986)

5.9.4 Lowering of cloud base

(a) during continuous *rain*:

  (i) Base will be at height where temperature lapse changes from positive to less positive or to negative.

  (ii) Base will be at a height where wet bulb or dew point is a minimum.

  (iii) If rain is of sufficient duration a ceiling will occur below 2000 feet; most frequently it is a ceiling below 800 feet.

  (iv) During continuous rain a ceiling does not generally occur at the height of temperature discontinuity and/or maximum humidity until after the occurrence of a ceiling corresponding to the next higher level of temperature and/or maximum humidity.

  (v) Ceiling remains practically constant until the next lower cloud layer appears and increases in amount for its base to become the ceiling.

  Note: Ceiling is the height above ground of the base of lowest cloud layers covering more than half the sky. (Method devised for USA).

(b) In intense *showers*: the cloud base may lower rapidly by 300 m (1000 ft) or more, rising rapidly again as the shower moves on.

(c) *Falling snow*: tends to establish an isothermal lapse rate just beneath the initial 0 °C level (see 5.2.2, Fig. 5.1).

Findeisen (1940)
Goldman (1951)
HWF (1975), Chapter 19.6.2

5.9.5 Seeder and feeder clouds

When a very moist low-level air flow is forced to rise over high ground, thick stratus/stratocumulus layers form. By itself this cloud may produce rather little precipitation, especially if the wind speed is strong and there is insufficient time for precipitation to form before the air flows down the lee side of the hills. If, however, rain is already falling from upper-level cloud layers (seeder clouds) above the stratus/stratocumulus, it will fall through the low-level (feeder) cloud, capturing the droplets within, considerably enhancing the rainfall rate at the surface.

The degree of enhancement depends on:

(i) seeder rate >0.5 mm h^{-1};

(ii) high θ_w in low-level air;

(iii) near-saturated air upwind;

(iv) strong low level wind ahead of cold front >30 m s^{-1} at 900 m for heaviest rain;

(v) favourable topography — not too steep.

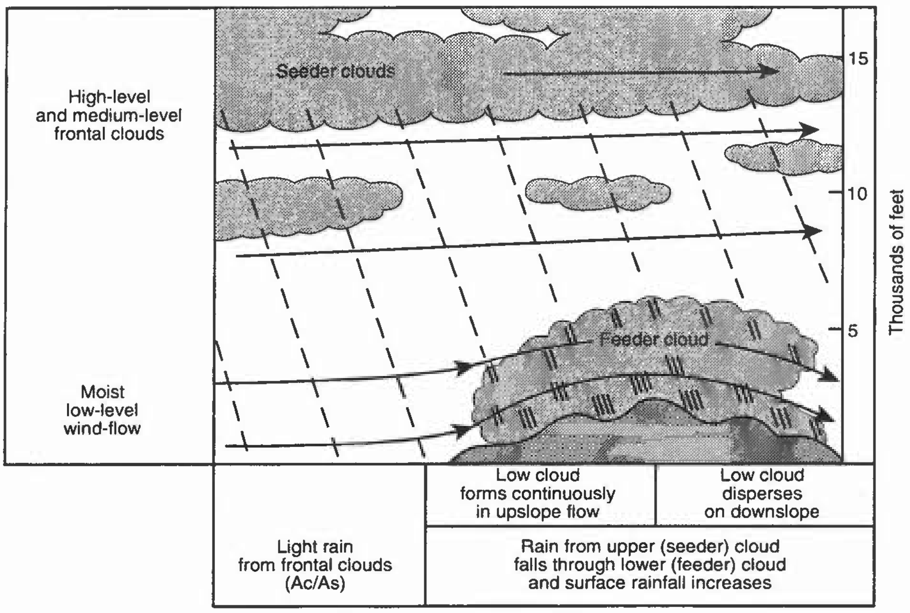

Figure 5.16. The seeder–feeder rainfall effect over hills.

This is a common mechanism for producing heavy rain over high ground in warm sectors approaching the UK from the south-west quadrant and is illustrated in **Fig. 5.16.**

Horizontal variations in surface rainfall intensity over high ground are on the same scale as the hills and valleys.

Curruthers & Choularton (1983) **Smith (1989)**
Robichaud & Austin (1988)

5.9.6 Mountain complications

A 'bad scenario' is as follows **(Fig. 5.17)**: a freezing level may exist in free air at about 1000 m. A warm front from the west, with low dew-point south-easterly air undercutting ahead of the front, gives cloud cover producing snow on hill tops down to 500 m. Below freezing level the melting snow lowers the air temperature and delays melting; as the freezing level descends, the snow level also lowers. Undercutting dry air also encourages further reduction in air temperature. It is not unusual for the freezing level to descend to the surface, with snow at all levels, resulting in widespread hill fog at sub-zero temperatures.

Local Weather Manual for Scotland (1994)

5.9.7 Freezing rain from elevated layers

(i) Rain at the surface at temperatures >0 °C is quite possible when there is freezing rain aloft; typically snow, falling through the melting level in a warm sector, encounters polar/arctic air beneath the warm/occluded front and re-freezes before melting again in the lower-level melting band. Freezing rain in such elevated layers from mid-latitude storms occurs about 1% of the time in the UK; oceanic freezing rain will not, generally, be a threat to high-flying transport aircraft.

(ii) Cumulonimbus clouds have sometimes been reported to have large water contents composed of precipitation-sized drops well below 0 °C.

Ahmed et al. (1993)

5.9.8 Severe low-level icing (rain ice)

(i) The forecaster must identify the contrast between low-level, dry pre-frontal air and moist air aloft. In the example **(Fig. 5.18(a))**, an occluding warm sector was pushing slowly east across England and France over an increasingly cold continental south to south-easterly airstream. Freezing rain became widespread over northern France.

(ii) The necessary detail will not be resolved from numerical forecast ascents **(Fig. 5.18(b))**; furthermore any ascent is a 'snapshot' and allowances must be made for this.

HWF (1975), Chapter 19.7.8

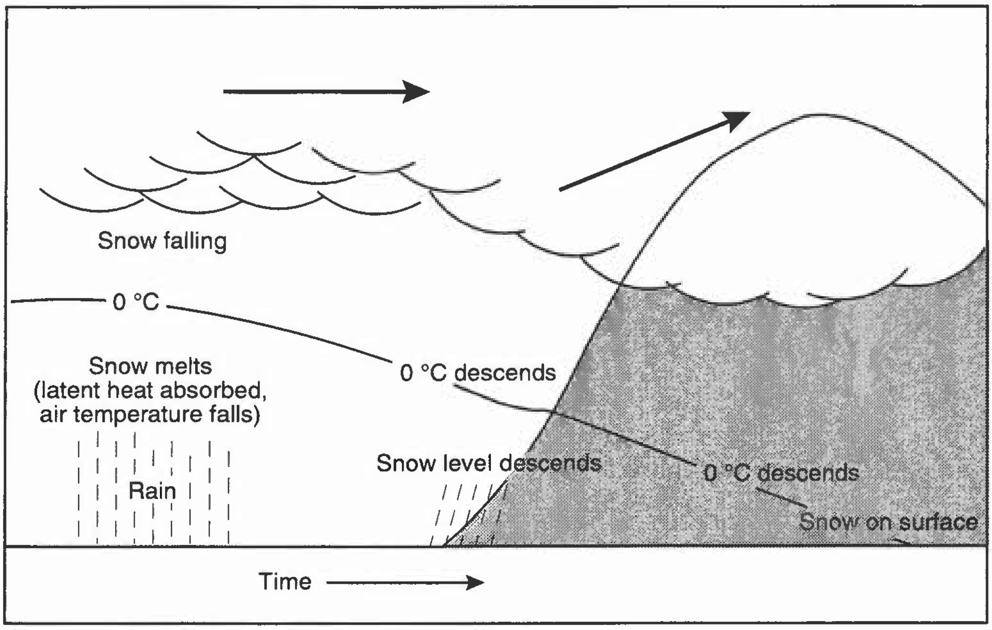

Figure 5.17. Mountain weather — the 'bad scenario' (see 5.9.6 for details).

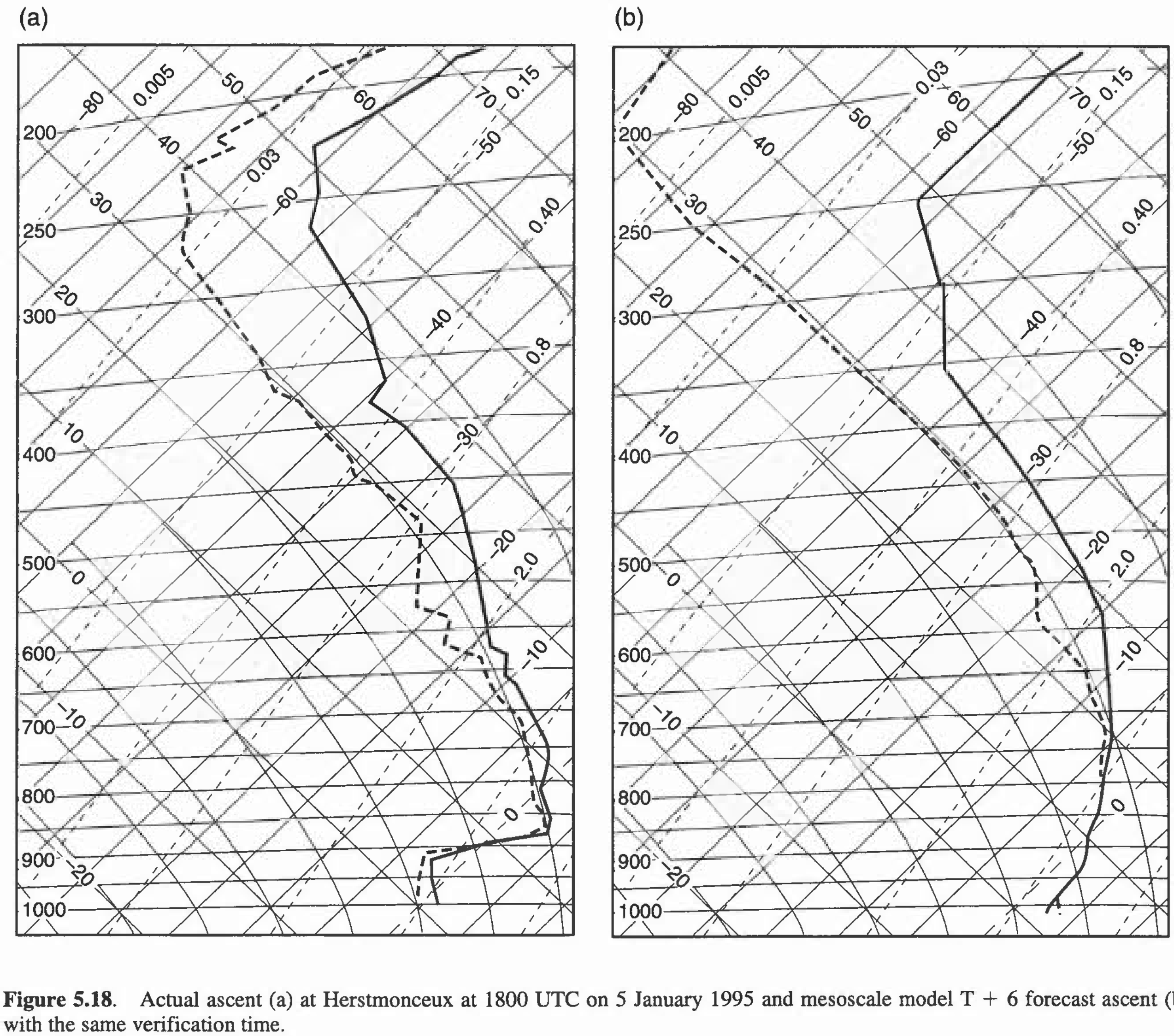

Figure 5.18. Actual ascent (a) at Herstmonceux at 1800 UTC on 5 January 1995 and mesoscale model T + 6 forecast ascent (b) with the same verification time.

5.10 Criteria for precipitation reaching the surface as snow or rain

5.10.1 Factors to consider:

(i) Site — high inland more prone than low coastal.
(ii) Orographic uplift can significantly lower the freezing level.
(iii) Snow cover — cover chills lowest layers implying that snow can be forecast with a slightly higher 1000–850 hPa thickness than would be expected.
(iv) Warm front approaching — if air preceding front is cold then precipitation can start as snow or freezing rain before turning to rain. A slow-moving warm front may develop a shallow wave and dramatically increase the potential for significant snow accumulations.
(v) In a tropical maritime air mass with wet-bulb temperatures of 10 °C or more, snow rarely falls within 50 miles of the front; if the 'warm' air mass is polar maritime then a wet-bulb temperature of about 4.5 °C in the cold air may give snow right up to the front.
(vi) Polar lows — note that NWP models cannot handle small-scale features well. Satellite and radar imagery will aid identification.
(vii) In winter a major problem is determining whether the precipitation will reach the ground as rain, snow, or rain and snow mixed; if no ice particles are present in a stratiform cloud, coalescence can still produce supercooled droplets that freeze on reaching the ground.

Various criteria have been derived for estimating the probability of precipitation in the United Kingdom falling as snow rather than rain. The following summary lists them in a rough order of merit; other techniques are presented.

Table 5.8.

		Percentage probability of snow				
Technique		90%	70%	50%	30%	10%
Adjusted value of 1000–850 hPa thickness (Boyden)	(gpm)	1281	1290	1293	1298	1303
Height of 0 °C isotherm (see Note (ii))	(hPa)	12	25	35	45	61
Surface temperature	(°C)	−0.3	1.2	1.6	2.3	3.9
1000–500 hPa thickness	(gpm)	5180	5238	5258	5292	5334

Lowndes et al. (1974)

5.10.1.1 Boyden's technique

The 1000–850 hPa thickness needs adjustment for the 1000 hPa height (H_{1000}) and the height of the ground above sea level (H_{GR}). The adjustment (m) is given by $(H_{1000} - H_{GR})/30$ and this quantity may be conveniently read off from **Fig. 5.19(a)**. Alternatively, the snow predictor nomogram, **Fig. 5.19(b)**, incorporates these corrections; it illustrates how a snow probability may be estimated.

Boyden (1964)

5.10.1.2 Height of 0 °C wet-bulb temperature technique

The height of the 0 °C wet-bulb temperature additionally takes into account latent cooling effects.

Table 5.9.

Height of 0 °C wet-bulb temperature	Form of precipitation
3000 ft or over	Almost always rain; snow rare
2000–3000 ft	Mostly rain; snow unlikely
1000–2000 f	Persistent rain readily turns to snow
Below 1000 ft	Mostly snow; only light or occasional precipitation falls as rain.

Beware of cold surface air undercutting warm; Hand uses mean temperature of the lowest 100 hPa (5.10.1.3).

HWF (1975), Chapter 19.7

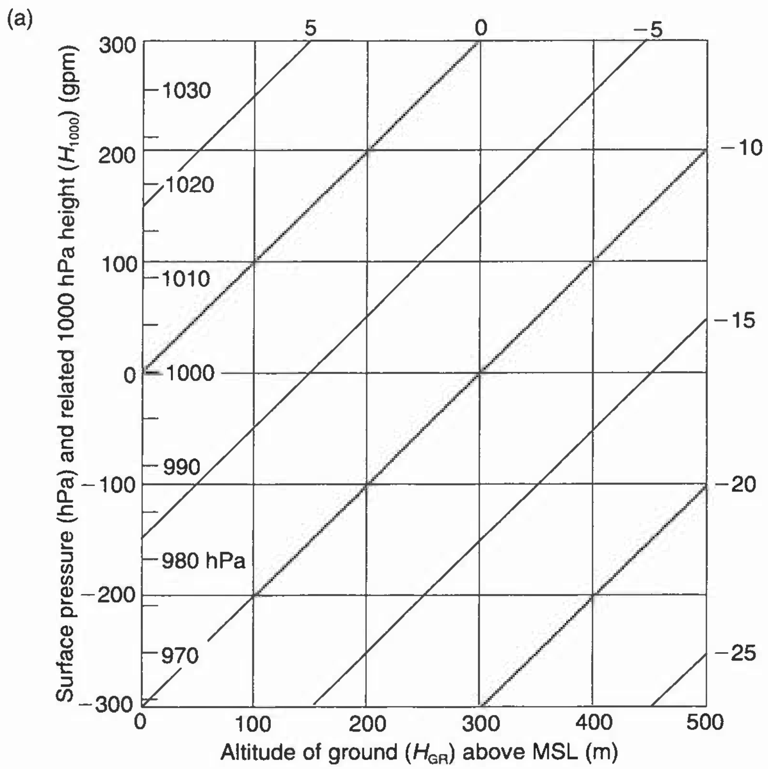

Figure 5.19(a). Forecasting the probability of precipitation as snow. This nomogram indicates the adjustment to be made to the 1000–850 hPa thickness to allow for the 1000 hPa height (or surface pressure) and altitude of ground above sea-level.

5.10.1.3 Hand's rule

Table 5.10. Use of mean temperature of lowest 100 hPa above ground to predict type of precipitation at the surface.

Mean temperature (°C) in lowest 100 hPa above surface	Precipitation type usually reaching surface
<−1.5	snow
−1.5 to 0.5	sleet
>0.5	rain

In heavy and persistent precipitation lowest layers will be further cooled by latent heat of evaporation.

Hand (1986)

5.10.1.4 Initial wet-bulb potential temperature level technique

Initial wet-bulb temperatures at which snow is likely for prolonged frontal precipitation are given in **Table 5.11**; the influence of frequent heavy showers/cold downdraughts can turn precipitation to snow with higher initial temperatures than from frontal precipitation.

Table 5.11. The relationship between downward penetration of snow beneath the 0 °C level and the initial wet-bulb temperature.

Type of precipitation	Initial wet-bulb temperature level (°C)	
	to which snow will descend	below which snow is unlikely
Prolonged frontal	+2.0 °C	+2.5 °C
Extensive moderate or heavy instability	+3.0 °C	+3.5 °C

HWF (1975), Chapter 19.7.6.1

(b)

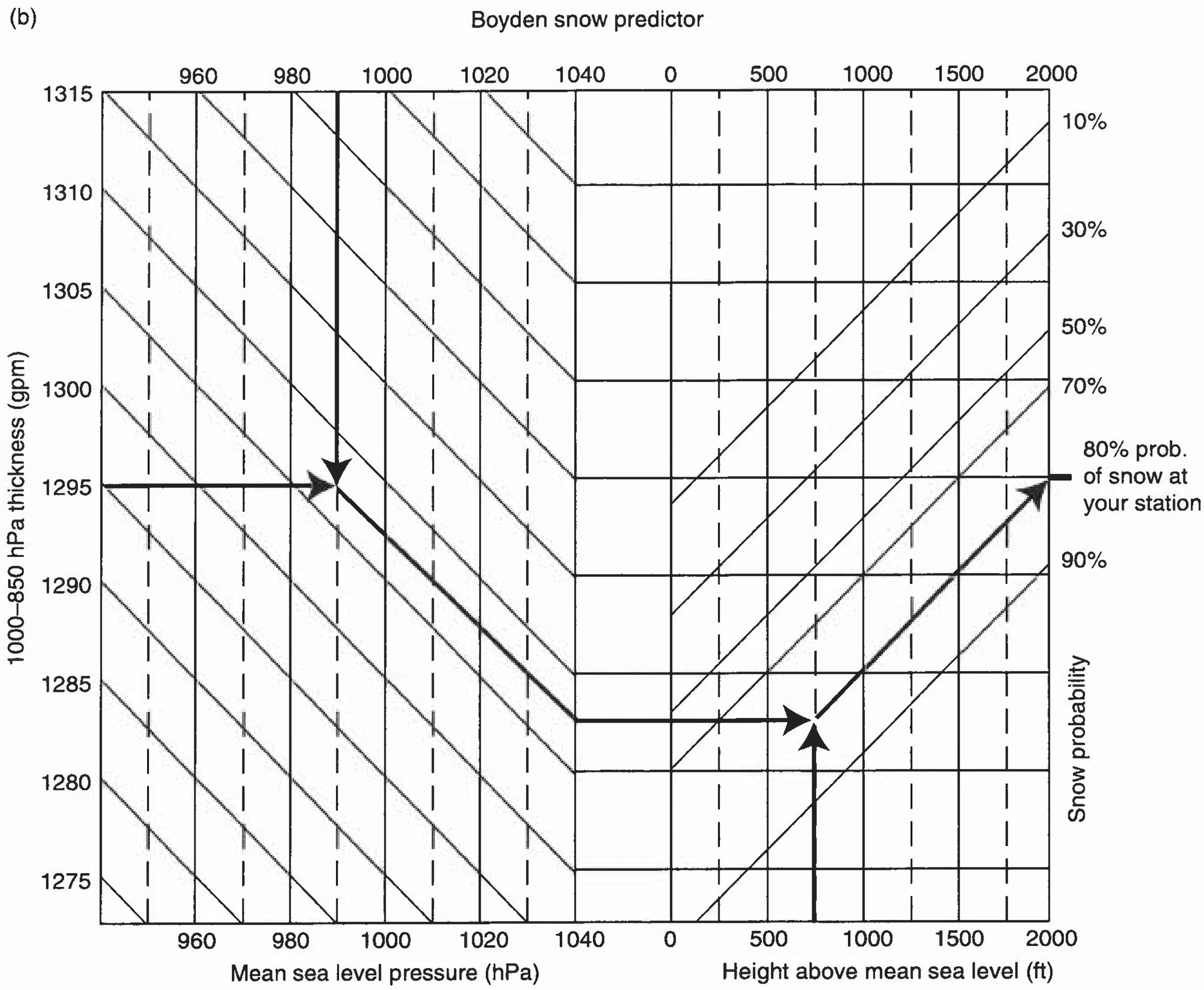

Figure 5.19(b). Forecasting the probability of precipitation as snow. How to find the probability of snow at a station with an elevation of 750 ft (230 m), a surface pressure of 990 hPa and a 1000–850 hPa thickness of 1295 gpm.

5.10.1.5 Screen wet-bulb temperature technique (Lumb)

For an exposed station at height H (in hundreds of metres) in central and western regions of the UK under moderate easterly, or stronger, winds:

(i) for elevations up to 170 m:
Rain turns to melting snow if: $T_w<(2.1 - 0.6H)$ °C
Rain turns to lying snow if: $T_w<(0.6H)$ °C
where T_w is surface wet-bulb temperature when precipitation begins.

(ii) for elevations 170 to 350 m:
Snow probable if: $T_w<(2.1 - 0.6H)$ °C.

(iii) If $T_w>2.5$ °C rain is more likely than sleet, irrespective of elevation.

Winds should be at least moderate with a good cover of low- or medium-level cloud.

The method is summarized in **Fig. 5.20**.

Lumb (1986)

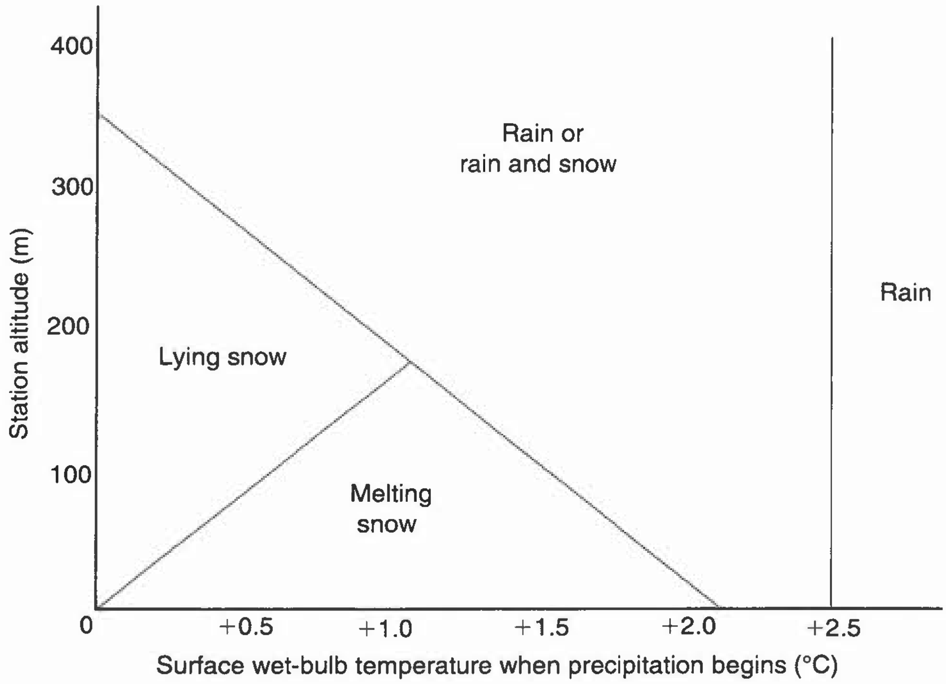

Figure 5.20. Diagrammatic form of Lumb's method for snow prediction.

5.10.1.6 Booth's snow predictor

A well tried snow predictor index, I_s, is based on the relationship: $T_w \approx (T + T_d)/2$ and I_s is defined as: $T + T_d$. The predictor can be plotted on a surface chart to show the most likely areas of rain, snow or rain turning to snow according to the **Table 5.12**:

Table 5.12.

Precipitation expected		I_s
Light intensity:	(i) probably rain if:	≥ 2
	(ii) probably snow if:	≤ 0
If continuous moderate or heavy precipitation is expected: (if there is no warm advection at or near the surface).		≤ 7

Rain may also turn to snow over areas into which colder air with $I_s \leq 7$ is advected.

Booth (1973)

5.10.1.7 Varley snow predictor

This predictor has proven skill in forecasting the level above which moderate or heavy snow (frontal or convective) will fall in the SW Midlands–Wales area. It is used for the lowest 2000 feet (610 m) and assumes:
$T_w \approx (T + T_d)/2$, lapse rate near the surface $\approx$ 2 °C per 1000 ft, and the snow will fall when T_w is 2 °C or below.
The height of the melting layer above the station in terms of T and T_d is:

$$250 \times (T + T_d - 4\ °C)\ \text{feet} \qquad [\text{or } 76 \times (T + T_d - 4\ °C)\ \text{metres}]$$

or, in terms of T and T_d: $[((T + T_d)/4) - 1]$ in thousands of feet
where T is the surface dry bulb, T_d the surface dew-point temperature and T_w the surface wet-bulb temperature. If, for example, $T + T_d$ in rain at a station at 20 feet is 8 °C, then snow in that area is likely at levels above about 1000 ft.

Table 5.13. An at-a-glance estimate of the height above which snow is likely in terms of T and T_d

$T + T_d$ (°C)	4	5	6	7	8	9	10	11	12
Melting height (ft)	surface	250	500	750	1000	1250	1500	1750	2000

5.11 Snow

5.11.1 Synoptic situations for snow in the United Kingdom

Over Britain, over 70% of substantial and extensive falls of snow are associated with a warm front or warm occlusion approaching from between south and west (cold fronts being less important). In such situations a large supply of moisture is available. Frontal progress is very slow as the warm air rises over a much colder continental easterly flow at low levels, so that snowfall is prolonged.

On 20% of occasions snow/blizzard conditions are due to polar lows which, in a northerly airstream, can produce substantial but more localized snowfalls — more frequently in Scotland than further south.

Easterly winds bring snow showers, intensified by the passage of troughs, and give substantial falls of snow in eastern Scotland and England. Exposed north-facing coasts as far south as Norfolk and Kent can be seriously affected when there is an unstable northerly airflow in winter.

Fig. 5.21 illustrate the annual incidence of days with (a) snow falling, and (b) lying.

Chandler & Gregory (1976)
Wild et al. (1996)

5.11.2 Lying snow

Large amounts of lying snow will greatly distort temperature-level structure, lowering the natural condensation level, the cloud base and increasing hill fog.

Local Weather Manual — Scotland (1994)

5.11.3 Snow over high ground

The 80%, 50% and 20% snow probability lines on numerical model forecast charts are based on forecast 1000–850 hPa thickness values and refer to mean sea level. The graph **(Fig. 5.22)** has been prepared to help forecasters to adjust these values in order to estimate the probability of snow at higher levels. Thus a snow probability forecast of 20% at mean sea level from the grid-point output of a numerical model becomes 50% at 500 ft (150 m) (e.g. most of the Marlborough Downs and the Cotswolds) and over 70% at 1000 ft (300 m).

5.11.4 Drifting of snow

With the temperature below 0 °C, drifting of dry/loose snow starts when the wind speed exceeds 12 kn. Serious drifting occurs with winds above 17 kn.

A *blizzard* is defined by the UKMO as 'the simultaneous occurrence of moderate or heavy snowfall with winds of at least force 7, causing drifting snow and reduction of visibility to 200 m or less'

Met. Glossary (1991)

5.11.5 Visibility in snow (see 3.10)

5.11.6 Thawing of snow

(i) Generally warm rain is the most effective agent for removing snow in winter since screen temperatures will not rise much above 0 °C over extensive areas of snow.
(ii) Insolation is most effective in other seasons, although it will have a negligible effect in hollows and north-facing slopes.
(iii) A depth of 150 mm (6 in) of snow requires either a continuous mild environment for several days or about 25 mm of rain to dispel it.
(iv) Since the advancing air will be cooled, thawing is less away from the windward edge of the snow cover.
(v) A screen temperature of 3 °C thaws 25 mm of snow in 24 hours, but if this warm air invasion is combined with appreciable rain, then 50 to 100 mm of snow thaws over the same period.

HWF (1975), Chapter 19.7.7

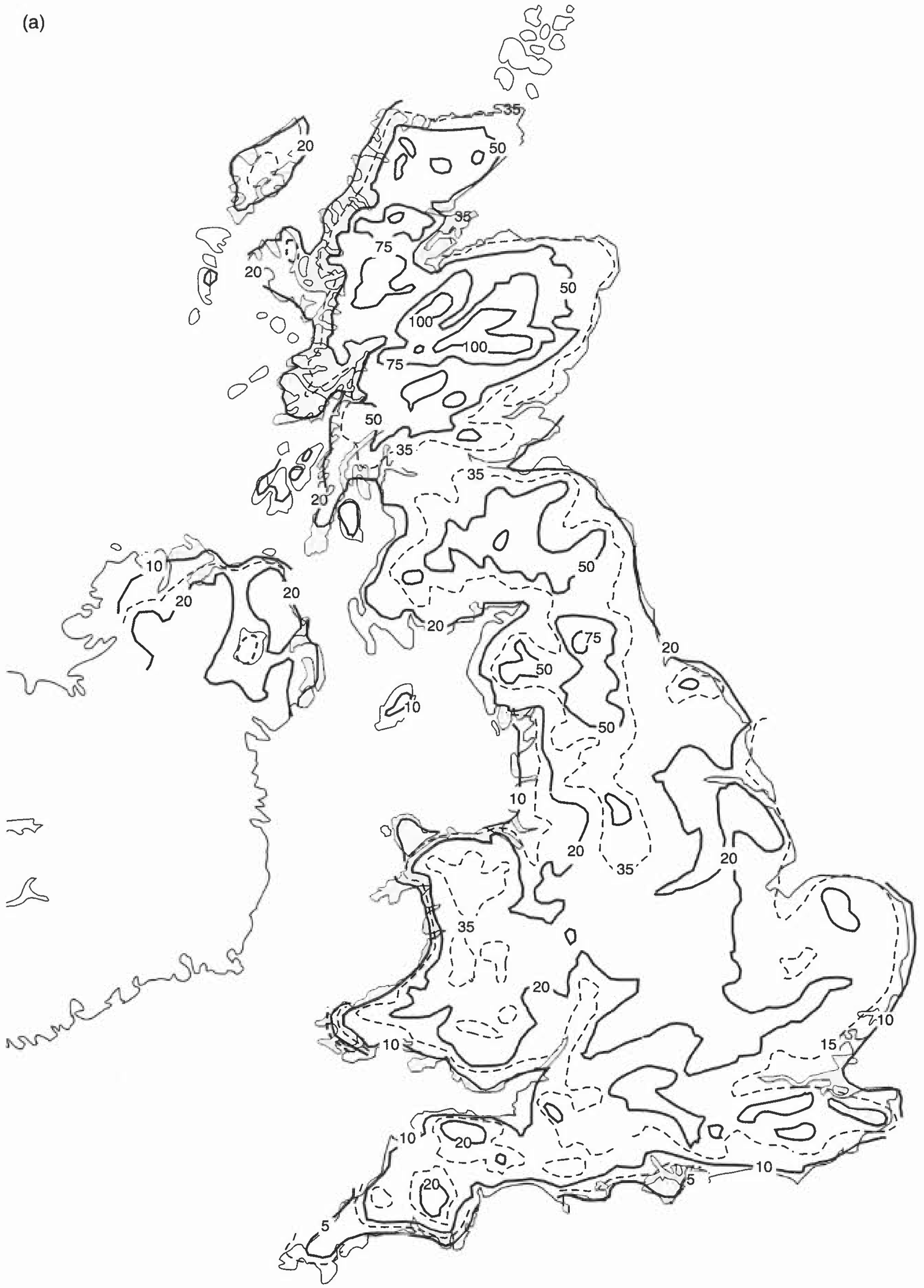

Figure 5.21(a). Annual incidence of days with snow falling.

(b)

Figure 5.21(b). Annual incidence of days with snow lying (October–May).

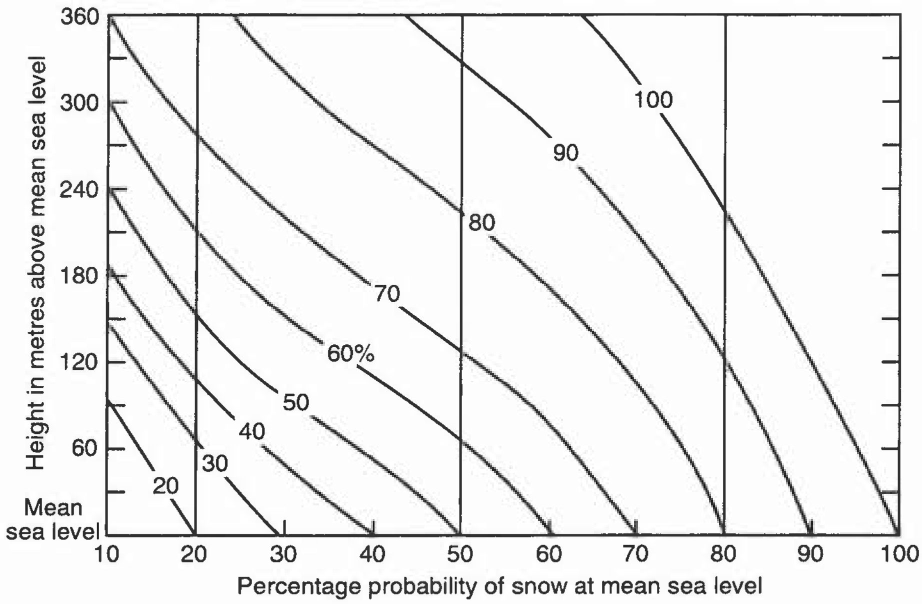

Figure 5.22. Snowfall over high ground. The increased probability of precipitation falling as snow over high ground, given its probability of occurrence at mean sea level.

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CHAPTER 5 — LAYER CLOUDS & PRECIPITATION

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