CHAPTER 4 — CONVECTION AND SHOWERS

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4.1 Forecasting convective cloud

The initial upward motion of an air parcel is provided by buoyancy from surface heating, convergence on a range of scales, mass ascent or orographic forcing. Condensation and entrainment mark the subsequent progress of the parcel upwards under various instability criteria until it achieves its upper limit. A simple model of the thermodynamic processes which occur can be represented on the tephigram which, however, shows only the static stability which does not always give a complete description of the true stability of moving air.

4.1.1 Instability definitions

Conditional describes a temperature sounding in a layer where the actual lapse rate lies between the SALR and the DALR.

Potential is shown by the decrease of wet-bulb potential temperature with height through a layer. This form of instability is released during mass ascent of air (e.g. over a hill, or in a rising frontal air current).

Latent an atmosphere that possesses potential instability may also possess latent instability (the converse is not necessarily true); a parcel of air with negative buoyancy may undergo forced ascent past its level of free convection. It then continues under its own buoyancy. The SALR through the wet-bulb temperature at any level on an ascent cuts the environmental temperature curve higher up. Latent instability describes, for example, a temperature sounding which is stable near the ground and unstable above, typically at night. Even though no convection is occurring at the time of the sounding, it may be released later in the day as a result of solar heating.

Met. Glossary (1991)

4.2 Constructions on a tephigram

4.2.1 Parcel method

Fig. 4.1 illustrates the general 'parcel' method, using a temperature sounding made before dawn (usually midnight) to forecast cumulus cloud during the day. ABCD is the environment curve. T and T_d are the surface temperature and dew point expected as a result of daytime heating. BU is the condensation level, CV is the level where the lapse-rate of the environment decreases to less than the SALR.

(i) This method assumes that a small parcel of air, of negligible mass, is warmed at the surface and becomes buoyant, rising up through the environment without disturbing it or mixing with it.

(ii) In **Fig. 4.1** the path curve of the rising parcel is shown by TUVWX. At T the parcel is warmer than the environment at that level (A) and it rises. From T to U its temperature falls at the DALR. U is the condensation level and cloud base, at which point it is still buoyant and continues to rise, but cooling at the SALR. At W (Parcel Tops) the rising parcel has a temperature equal to that of the environment and it is no longer buoyant, but theoretically, it will continue to rise to X because of momentum gained during ascent.

(iii) Some tops may reach W and, in extreme cases, vigorous convection cloud may penetrate to X, which is defined as the level which is such that the 'negative area' XDWX equals the 'positive area' WVTACW. This latter area defines the convectively available potential energy, CAPE.

(v) For practical use a modification to this method assumes that entrainment, detrainment and frictional drag slow the ascent. As a result the ascent ceases at a lower level than that predicted by the parcel method. The level C in **Fig. 4.1**, where the lapse-rate of the environment becomes less than the SALR, is the cloud-top level for most cumulus (Slice Tops).

Summary:

U is Normand's point

V is taken as the tops of most of the clouds

W is forecast for the tops of occasional large clouds

X is forecast only when conditions seem favourable for exceptionally vigorous and deep convection (e.g. θ_w lapse rate ≥ 0).

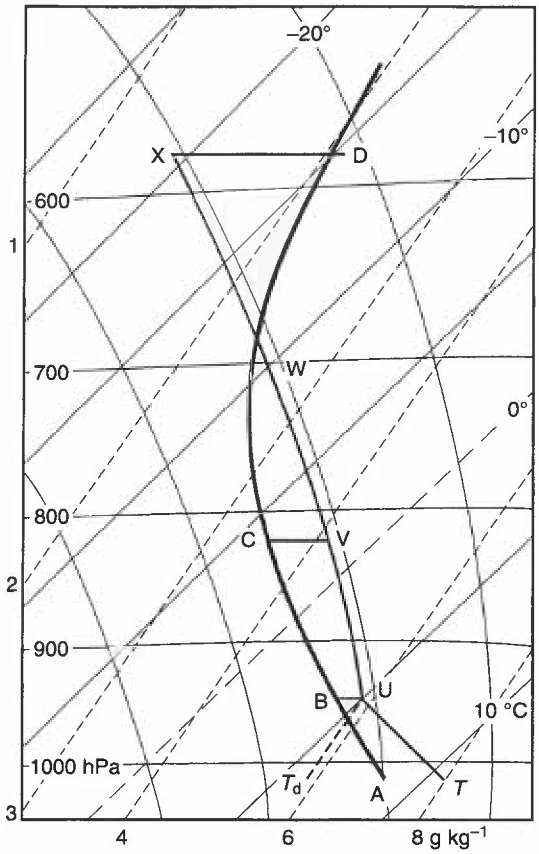

Figure 4.1. Tephigram construction for estimating the base and tops of convection cloud. See text for method of use.

4.2.2 Forecasting the cloud base of cumulus

4.2.2.1 Estimating the condensation level

The surface temperature (T) and dew-point (T_d) may be used to estimate the condensation level (H).

As a rough approximation:

$$H = 4\,(T - T_d)$$
$$H_m = 1.22\,(T - T_d)$$

where T, T_d are in degrees Celsius and H in hundreds of feet (H_m in hundreds of metres).

On a tephigram, Normand's theorem gives the condensation level at the intersection of a dry adiabat through T and a humidity mixing ratio (HMR) line through T_d. This is the level BU in **Fig. 4.1.**

4.2.2.2 Relationship between condensation level and cloud base

(i) While surface temperatures are rising quickly during the morning, the base of cumulus cloud is close to the calculated condensation level.

(ii) During the afternoon, when convective upcurrents are strongest, the base of cumulus may be up to 700 ft (25 hPa) above the condensation level as the moisture flux from the surface causes the mixing ratio to decrease in the lowest tens of metres near the ground.

(iii) These relationships are based on data from gliding sites where light aircraft and sailplanes provided frequent reports of cloud base. Aircraft observations from Bircham Newton and Aldergrove, made during the months from April to September, showed an average cloud base 25 hPa above the condensation level.

(iv) After the time of maximum temperature the cloud base remains almost constant although the falling surface temperatures imply a lowering of the condensation level.

(v) Even in an apparently homogeneous air mass there may be significant variations in the cumulus cloud base over short distances. The difference is usually greatest during the morning when the air is not well mixed. Different rates of heating over hilly and over low-lying areas can lead to temporary variations of the cloud base by as much as 2000 ft (600 m) within a distance of 1 km.

(vi) The daytime ground surface temperature and dew point of fairly extensive areas of elevated ground may not be greatly different from surrounding low-ground values; condensation levels may thus be at a similar height above elevated ground as condensation levels above adjacent lower ground — the 'plateau effect'.

4.2.3 Forecasting the heights of convective cloud tops

4.2.3.1 First estimates — using a tephigram

(i) An initial estimate of the cumulus/cumulonimbus cloud tops may be made following the method illustrated in **Fig. 4.1,** paying due regard to the limitations imposed on cloud-top height by entrainment etc. (the 'Slice Tops' approach). Observations often confirm the irregular level of actual cumulus tops.

(ii) It is suggested that, as a first estimate, the construction shown in **Fig. 4.1** is adopted.

(iii) Workstation satellite imagery can be used to see the spatial distribution of showers. By using temperature slicing of IR images an estimate of the cloud-tops temperature can be found. Then an estimate of cloud height may be made from an adjacent tephigram.

4.2.3.2 Choosing a representative ascent

(i) At the time a forecast is being prepared, the appropriate sounding will be upwind of the area and in the correct air mass. If there is cold advection, as indicated by wind backing with height, then the environment is likely to become more unstable with time.

(ii) Even if no fronts are expected to cross the area bringing major air-mass changes, there may be significant variations in the convective activity within the same air mass.

(iii) Enhanced convection may be associated with the occurrence of synoptic-scale features such as upper troughs or lows.

(iv) Similarly, suppressed convection may accompany an upper ridge.

(v) Mesoscale variations are also common, and may be associated with regions of low-level convergence/divergence or variations in low-level moisture content.

Local Weather Manuals (1994)

4.2.3.3 Prediction of surface dew points

(i) **Fig. 4.2** shows a tephigram construction for finding a representative daytime dew point from a night-time sounding.

(ii) The full line is the dry-bulb curve and the pecked line is the dew-point curve at dawn. The maximum surface dew point may be found by extrapolating down to the surface that part of the dew-point curve (DX) which has been unaffected by nocturnal cooling in the boundary layer. The dew point so obtained, A, is the maximum daytime value of the dew point. This will occur in the late morning when cumulus cloud first forms. The temperature T_{Cu}

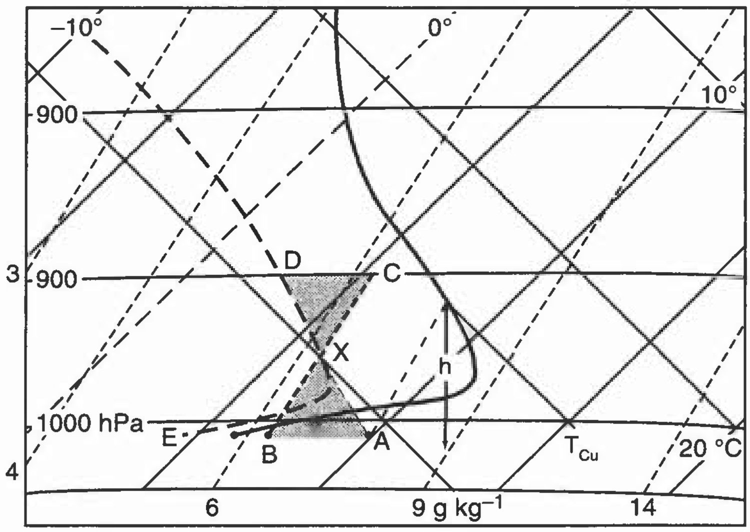

Figure 4.2. Tephigram construction for estimating a representative daytime value of the surface dew point, and temperature of first formation of cumulus. See text for method of use.

and height, h, at which the first Cu form is estimated by drawing up a constant HMR from A to the environment curve (for h) and then down to the surface along a DALR (for T_{Cu}).

(iii) With the onset of vigorous convection, at midday or soon after, the boundary layer develops an almost constant HMR. The afternoon value of the dew point (B) may be found by drawing a line of constant HMR (CXB) such that the total moisture content represented by the values along DX equals that represented by the values along XA. Since the HMR scale is not linear, DX should be slightly longer than XA. In practice it may be assumed that DX equals XA to a first approximation.

(iv) In this example it has been assumed that the mixed layer extends up to 900 hPa. In midsummer it may be necessary to raise this level to a point 150 hPa above the surface.

(v) On days with strong heating the reported dew points may show wide differences between adjacent inland stations by mid afternoon, especially if winds are light. Stations on or near the coast report much higher dew points when there is an influx of air from the sea.

4.3 Forecasting considerations

4.3.1 Synoptic-scale indicators of enhanced or suppressed convection

Note that two factors are sought — changing stability and a trigger mechanism. The forecast height of convective cloud tops suggested in **4.2.3** should be modified according to the synoptic-scale and mesoscale developments expected during the day. Such developments are, for example:

(a) Synoptic-scale ascent makes the air mass less stable. Thus, factors indicating an increased depth of convection are:
- (i) Approach of a trough, or low, in the 500, 300 or 200 hPa contour pattern (i.e. positive vorticity advection); convection decreasing rapidly after its passage.
- (ii) Advection of warm, moist, air at low levels.
- (iii) Advection of cold air at medium levels: the approach of a 1000–500 hPa thickness trough or ridge moving away.
- (iv) A trough, or increased cyclonic curvature, on the surface chart, or at time of maximum heating.
- (v) A convergence zone (e.g. due to a sea-breeze front, topography, or cold air drainage converging over a warmer sea surface, as in the Bristol Channel area).
- (vi) Over mountains in summer when the winds are light (4.6).
- (vii) Airflow over the sea across the isotherms towards warmer water.

(b) Synoptic-scale descent makes the air mass more stable. Thus, factors indicating a decreased depth of convection are:
- (i) Approach of a ridge in the 500, 300 or 200 hPa contour pattern (i.e. negative vorticity advection).
- (ii) Advection of dry air at low levels, tending to evaporate cloud.
- (iii) Advection of warm air at medium levels: 1000–500 hPa trough moving away or ridge approaching.
- (iv) Increasing anticyclonic curvature, or the approach of a ridge, on the surface chart. (Inversions will inhibit convection even if parcel stays to right of environment curve.)
- (v) Strong wind shear will often inhibit vertical extent of convection; however, marked shear is necessary for supercell storms.
- (vi) At the left entrance or right exit of a jet stream.
- (vii) Over valleys or lower ground surrounded by mountains when upper winds are light in summer.
- (viii) Airflow over the sea across the isotherms towards cooler water.

Note: continued convection will gradually stabilize the air due to the adiabatic warming by subsidence between the clouds, and transport upwards of latent heat from the surface (but not if synoptic-scale ascent is taking place).

Differential heating between wet and drier ground, or vegetation differences, can result in the development of mesoscale thermal gradients and corresponding circulations, and convergence zones.

Bader et al. (1995), Chapter 6

4.3.1.1 Summary of forecasting pointers — convective cloud

Decide which air mass will affect the station. Look at its history: did showers develop in it yesterday? If so:

Where? Over sea/coasts/hills/inland?
When? Throughout 24 hours or only at the time of maximum temperature?
At what temperature?
What factors since yesterday have changed the stability?
Heating/cooling from below — advection over warm sea, etc.
Warming/cooling aloft — warm or cold advection.
Increase/decrease in moisture content — by evaporation from a surface/advection of moister air aloft.
Wind direction change may have brought air with markedly different low-level characteristics into the local area, or may do so later in the day.

From representative ascents: what factors might release potential/latent instability?

Low-level convergence due to:
(i) Cyclonic curvature of isobars.
(ii) Along sea-breeze front or where two sea-breeze fronts meet.
(iii) Falling surface pressure.
(iv) Coastal convergence.

Forced mass ascent due to:
(i) Orographic uplift.
(ii) Divergence aloft — upper troughs, etc.
(iii) Along a front.
(iv) Intense surface heating — over a plateau.

HWF (1975), Chapter 19.7

4.3.2 Organization of shallow convection

4.3.2.1 Convection and waves (see 1.3.2.4 and 10.3.1.1)

Thermal activity can be modified by the gravity waves that have, in turn, been generated by the convective activity.

Down-wind patterns (streets)
If the depth of convection is limited by a stable layer and cloud streets form, the waves form parallel to the streets when the winds (usually constant in direction with moderate speed increasing slightly with height) blow across them.

Cross-wind patterns (waves)
If there is no change of wind direction with height and convection increases to 3 to 4 km, transverse waves may form at right angles to the cloud streets. Such waves are influenced by topography but can form even when strong convection extends far above any mountain summits.

Cumulus development will be enhanced in the ascent regions of the waves, and suppressed in the descent regions; the resulting cloud pattern is unlikely to be very well defined.

Booth (1980)
Bradbury (1990)
Ludlam (1980)

4.3.2.2 Sea breezes and other convergence zones (see 1.3.1.1 and 4.6)

4.3.2.3 Convection over the sea (see 10.3.1.2)

4.3.3 Forecasting thermals for glider flights

The best gliding conditions usually occur when the top of the convective layer is marked by an inversion between 5000 and 8000 ft, and thermals are marked by shallow Cu.

Thermal strengths are classified as nil, weak, moderate or strong as follows:

Table 4.1.

Thermal category	Max. rate of climb (kn)	(m s^{-1})	Cu bases at, or dry adiabatic conditions to (feet agl)
Nil	0	0	<2000
Weak	≤3	≤1.5	≥2000<3000
Moderate	>3≤6	>1.5≤3	≥3000<5000
Strong	>6	>3	≥5000

Mean rate of ascent will be less.

Fig. 4.3 illustrates maximum lift (a) in cloudless thermals, and (b) in cumulus clouds, the latter deduced from the tracking of free balloons by radar. Thermal activity begins to decrease about an hour after maximum surface temperature has been reached. After 1700 UTC thermals usually subside quickly; lift may still be found over towns and south-west facing slopes well into the evening.

An empirical method for estimating thermal strength, based on numerous French glider pilot reports, is illustrated in **Fig. 4.4.** The prediction requires an estimate of cloud base:
(i) Move horizontally across from an estimate of cloud base (e.g. 5500 ft) to one of the diagonal line labelled with the cloud amount (e.g. 1/8).
(ii) Follow line from intersection vertically to the Average Lift scale (e.g. 4.25 kn).
(iii) Continue down to intersect the diagonal lie and read off the Max Lift (e.g. 7 kn).

Summary: a prediction of 1/8 cloud at 5500 ft gives a forecast average lift of 4.25 kn, peaking at about 7 kn.

Lee waves enable gliding to high altitudes in the lee of quite moderate hills; vertical velocities are generally between 5 and 10 kn (2.5 to 5 m s^{-1}), exceptionally speeds >25 kn (12.5 m s^{-1}) and heights in excess of 30,000 ft have been recorded. Often associated with such systems, however, are flight hazards such as rotors and severe downslope winds (1.3.3.4). Waves above isolated cumulus, cumulus streets and waves, enhanced by cumulus over mountains, offer good opportunities, for example, for cross-country flights.

Local Weather Manual for Southern England (1994)

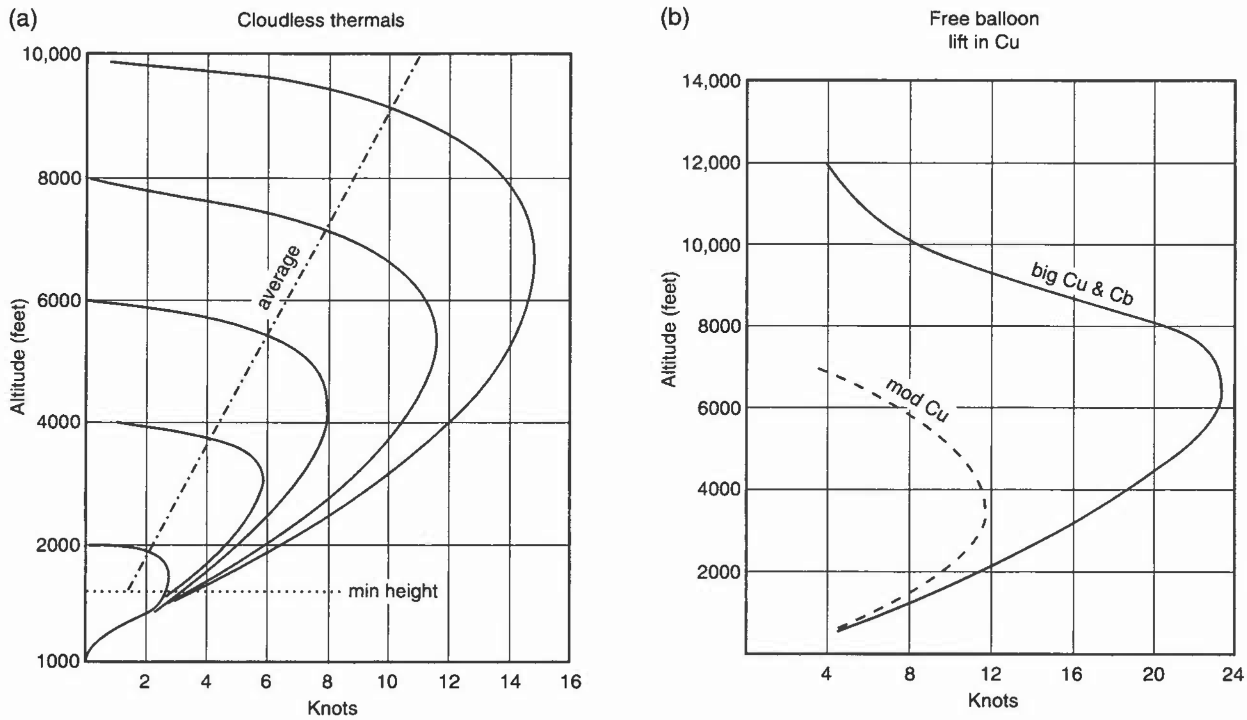

Figure 4.3. (a) Lift in cloudless thermals, and (b) lift in cumulus cloud.

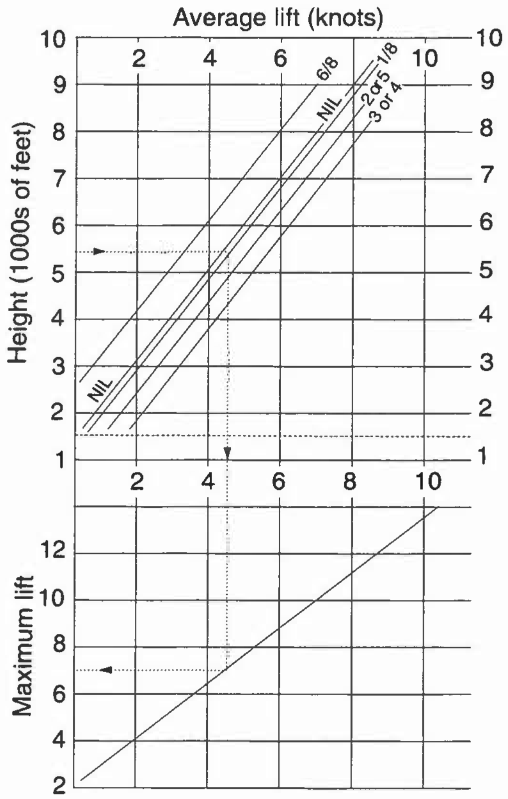

Figure 4.4. Empirically derived average and maximum thermal lift as a function of cloud base and cloud cover (see text for details).

4.3.3.1 Cross-country flights

Prediction diagrams of air trajectory, isobaric curvature, 850 hPa wind speed/direction, and potential temperature/dew-point depression data (due to Bradbury) may be 'scored' to assess long-distance prospects. **Table 4.2** summarizes favourable conditions over closed-circuit routes.

Table 4.2. Long cross-country flights over closed-circuit routes: favourable conditions

Previous air trajectory	From NW, N or NE (never from S).
Curvature of isobars	Anticyclonic.
Mean sea-level pressure	1023 hPa ($\pm$7 hPa).
850 hPa wind	Speed not more than 16 kn, direction between WNW and ENE through N.
Stability	Potential temperature decreasing about 3 °C between surface and 850 hPa at the time of maximum temperature. Depth of instability restricted to a stable layer below 700 hPa to prevent any shower activity.
Surface dew-point depression	11 to 18 °C by mid-afternoon.
Surface moisture and rainfall	State of ground dry at 06 UTC, no overnight rainfall.
Sunshine	At least 8 hours bright sunshine.
Visibility	More than 20 km.

Booth (1978)
Bradbury (1978, 1991a, 1991b)
Met O 6 Gliding Notes (1989)

4.4 The spreading out of cumulus into a layer of stratocumulus

4.4.1 Cloud cover beneath an inversion

When the depth of convection is limited by an inversion, cumulus tops may spread out to form an almost unbroken sheet which covers large areas and persists for long periods. This is common in subsided polar maritime air masses, particularly on the eastern flanks of anticyclones. **Fig. 4.5** illustrates an empirical method for estimating the cloud cover beneath an inversion. B is the condensation level, derived from the expected surface temperature, T, and dew point, T_d, BC is a saturated adiabat from cloud base to cloud top. DE is a dry adiabat from the base of the capping inversion, cutting BC at E.

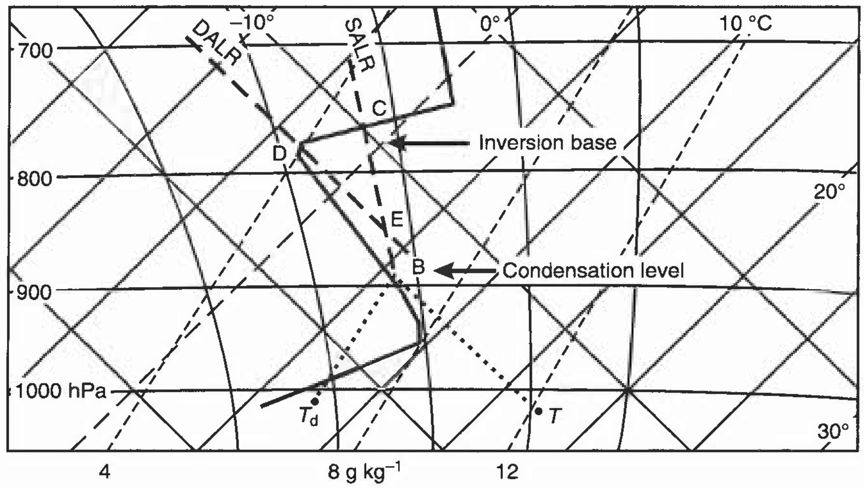

Figure 4.5. Estimating the spread out of cumulus cover beneath an inversion. See text for method of construction.

The expected cloud amount equals CE divided by CB, where the depths are conveniently measured in hectopascals and the answer is a fraction which can be expressed in terms of oktas.

4.4.2 Criteria for development of stratocumulus spread-out (see 5.8)

(i) An inversion or well-marked stable layer strong enough to halt all convective upcurrents even at the time of maximum insolation.
(ii) A lapse-rate close to the DALR from the surface to near the base of the inversion when convection starts.
(iii) Condensation level at least 2000 ft below the level of the inversion.
(iv) A dew-point depression of 5 °C or less in the layer between the condensation level and the base of the inversion.

4.4.3 Criteria for break-up of cloud sheet

(i) Decreasing surface dew points, lifting the condensation level to within 30 hPa of the inversion.
(ii) Increasing surface temperatures, sufficient to lift the condensation level to within 30 hPa of the inversion.
(iii) Continued subsidence, bringing the inversion down to within 30 hPa of the condensation level.
(iv) A weakening of the inversion, allowing cumulus tops to break through to a higher level.
(v) If the cloud layer is formed due to diurnal heating over land, nocturnal cooling usually results in the dispersal of the layer.
(vi) Cloud formed by convection over the sea shows no such diurnal variation.

HWF (1975)

4.5 Forecasting showers

4.5.1 Precipitation processes within continental and maritime clouds

(i) Concentrations of cloud condensation nuclei (CCN) decline over land by a factor of about five between the surface and 5 km while they remain fairly constant in maritime air where the average cloud droplet size is larger.
(ii) These factors have an important influence on precipitation development; precipitation may fall from a warm cloud of maritime origin with limited depth, due to the more efficient coalescence process with larger droplet size, while a similar cloud over land will not rain.
(iii) Air masses quickly acquire the CCN characteristics of the surface over which they are passing; for example, aerosol collected over the Irish Sea in a polar maritime north-westerly were found to be predominantly of a continental type.
(iv) Clouds with tops between 0 and −4 °C generally consist entirely of water drops; it is within this temperature range that the worst aircraft icing conditions are most likely.

4.5.2 Depth of cloud needed for showers

(i) The diagram at **Fig. 5.15** may be used as a rough guide to the intensity of both convective and non-convective precipitation, even though cloud-top temperature is not considered.
(ii) Showers are likely to be heavier in intensity than layer-cloud precipitation for the same thickness of cloud.

(iii) The diagram only applies if the difference in water content at the base and top of a shower cloud exceeds 1.5 g kg^{-1}.
(iv) Moderate Cu of maritime origin may give more precipitation than indicated.
(v) If the cloud depth is <5000 ft (1500 m) with cloud-top temperature warmer than −12 °C there is a 10% probability of showers; the figure rises to 90% for cloud depth >10,000 ft (3000 m). Exceptions to this rule are the 'drizzly' showers from warm maritime clouds and the shallow winter maritime/coastal showers common in northern areas in northerly polar flows with limited instability.
(vi) It is important that the top should certainly be <−4 °C, and generally <−10 °C, for any likelihood of showers by the Bergeron–Findeisen mechanism.

See 4.7.1.2 for guidance on the probability of thunderstorms related to the height of cumulonimbus tops.

Pettersen et al. (1945)

4.5.2.1 Cloud cover and lifetime

(i) Cover — if RH at cloud level is 50% suggest 2 oktas;
if 75% suggest 5 oktas.
(ii) Surface observations routinely give larger amounts of convective cloud than are seen from the air or satellite, since the gaps between distant Cu are obscured from the ground-based observer by adjacent Cu.

Table 4.3. Depths, updraughts and lifetimes of Cu and Cb

Cloud type	depth (ft)	updraught (m s^{-1})	lifetime
Small Cu	1500	1–5	20 min
Large Cu	6000–15,000	5–10	1 hour
Cb	15,000 upwards	10–20	>1 hour
Supercell Cb	15,000 upwards	>50	>>1hour

4.5.2.2 Intensities of showery precipitation (UK Met. Office definitions)

Table 4.4.

Rain showers	Intensity (mm h^{-1})
Slight	<2.0
Moderate	2.0 to 10.0
Heavy	10.0 to 50.0
Violent	>50.0

Meteorological Glossary (1991)
Observer's Handbook (1982)

4.5.3 Showers and wind shear

The wind shear in the cloudy convective layer is a useful key to the persistence of individual showers:

(i) *No change in wind speed or direction* — there is mutual interference between coincident updraughts and downdraughts. This occurs with shallow clouds. Showers are very light and last only a few minutes.
(ii) *Increasing wind speed but little change in direction with height* — the clouds slope forwards; thus downdraughts fall into the inflowing surface air, cutting off the updraughts and reducing the lifetime of the showers. This commonly occurs in maritime airstreams with moderately deep convection and light/moderate showers. Though each shower may have a life-span of 20–30 minutes, specific places are affected for a shorter time.
(iii) *Strong vertical wind shear* — usually associated with strong isobaric temperature gradients and an upper stable layer. Not compatible with showers (although intrinsic to severe storm development — see 4.7.7.2).

HWF (1975), Chapter 19.7.3.3 **Ludlam (1980)**

4.6 Topographically related convection

Satellite and radar imagery confirm that the distribution of convection is rarely random. Cloud bands and precipitation repeatedly occur in similar air masses relative to the same topographic features (**Figs 4.6 and 4.7** — radar data were not available for Scotland at the time). Forecasters may be familiar with other local patterns.
Topographically induced cloud forms as a result of:
(i) convection generated in unstable air over relatively warm surface;
(ii) thermally driven land or sea breeze;
(iii) frictional convergence induced by coasts (cloud bands giving intense showers inland can result from a winter airstream with a long fetch down the North Channel);
(iv) deflection of airstream by hills or headlands.

Isolating the particular cause may not be possible but satellite imagery may provide some clues, e.g. (i) and (ii) will be seasonal and diurnal, (iii) will depend on wind direction (10.3).

(a) Bands of enhanced convection (either in an otherwise cloud-free area or in an area of abundant showers) are most common in polar maritime air masses (wind directions between south-west and north).
(b) Location of convective band maxima for airstreams from four directions are shown in **Figs 4.8 and 4.9**; they provide a fairly comprehensive picture of favoured locations and downwind penetration for cloud.
(c) When winds exceed 20 kn on windward coasts, dominant cloud areas are over high ground; such strong winds also disrupt the summer cloud pattern, correlated with the coast line, because the sea breeze cannot become established (double-headed arrows indicate that bands occur within a range of wind directions; single-headed arrows denote where successive areas of activity follow similar paths).
(d) Note that shallow summer convection often produces a pattern of Cu clouds which imitates the shape of the topography, but is displaced by 10 to 20 n mile (18.5 to 37 km) downwind.
(e) When off-shore winds are light, sea breezes may develop, enhancing the Cu just inland from the coast and delaying the downwind drift of Cu.

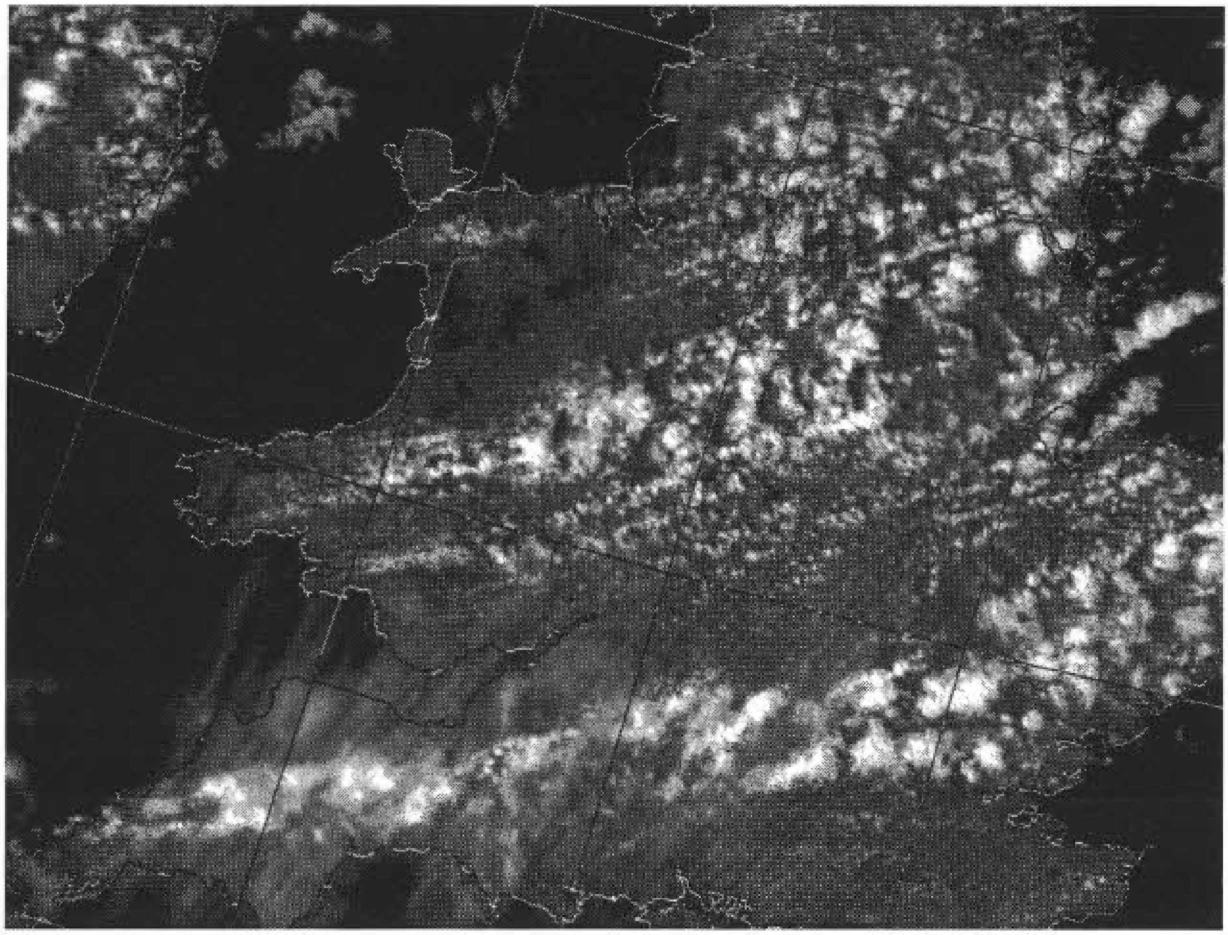

Figure 4.6. NOAA-9 visible image at 1406 UTC on 13 May 1986. (Photograph by courtesy of University of Dundee.)

(f) The approach allows a general forecast of 'sunny (or clear) periods and showers' to be more specific, with emphasis on the distribution of cloud and precipitation.
(g) Severe weather may develop if the steering level of convection lies along a convergence zone.

Bader et al. (1995), Chapter 6
Browning et al. (1985)
Hill (1983)
Monk (1987)
Orographic Processes (1993)

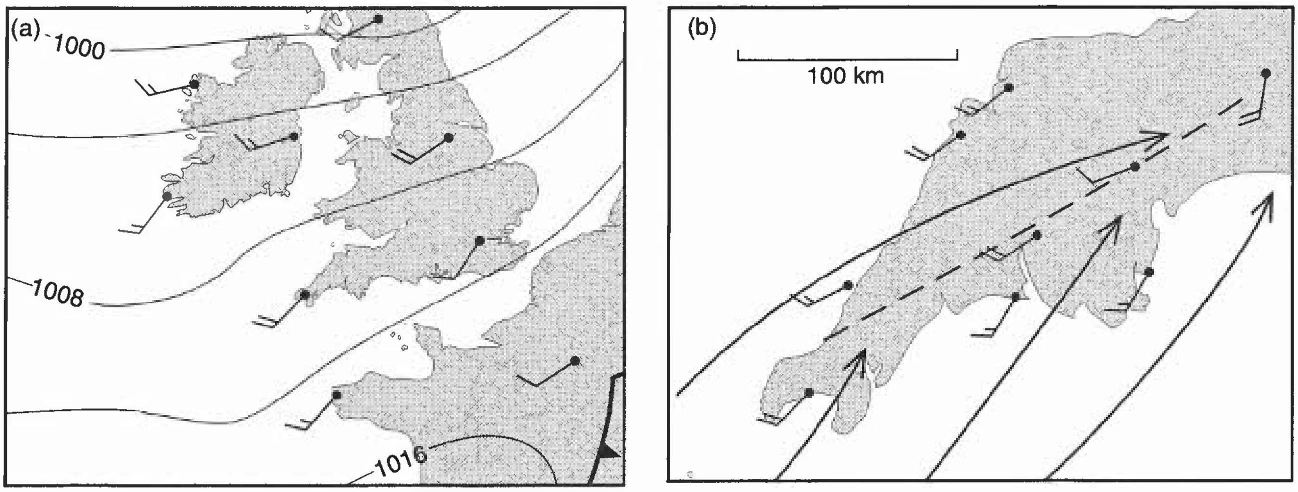

Figure 4.7. Surface analyses at 1400 UTC on 13 May 1986. (a) Isobaric analysis, and (b) available wind observations and streamlines over south-west England. The convergence line suggested by the cloud and precipitation is shown by the dashed line.

4.7 Forecasting cumulonimbus and thunderstorms

4.7.1 Main factors

At around −20 °C a significant proportion of cloud particles in an air parcel will be composed of ice crystals, giving the cloud boundary a fibrous appearance. This marks the transition of the cloud from large cumulus to cumulonimbus. The vertical wind structure will determine the lifetime and severity of the storm and whether it exists as a single cell or develops as a multicell storm.

4.7.1.1 Movement of thunderstorms: the steering level

With a cumulonimbus extending through a deep layer in which there is marked wind shear, the storm cloud is steered by the wind at the level approximately one third of its depth, measured from the base of the cloud, i.e.

$$H_{\text{base}} + 1/3\,(H_{\text{top}} - H_{\text{base}}).$$

In the United Kingdom, the steering level is often around 700 hPa. However, for a storm with a 5000 ft (1500 m) base and top at 40,000 ft (12,000 m) this gives a steering level of 16,600 ft (5000 m; 550 hPa).

Ludlam (1980)

4.7.1.2 Depth of cumulonimbus giving thunder

A useful guide is given by:

Table 4.5.

If cumulonimbus tops are at:	
<13,000 ft	— thunder unlikely
14,000–18,000 ft	— thunder probable
>18,000 ft	— thunder highly probable

A better guide may be: cloud-top temperature ≤−18 °C.

HWF (1975), Chapter 19.7.5

Frequently observed cloud bands

Infrequently observed cloud bands

Dominant cloud areas when wind >20 kn

L Cloud areas observed when wind <10kn

Cloud-free areas due to sea-breeze penetration

Coastal regions enclosed by thicker lines are cloud free due to sea-breeze penetration. With the stronger wind the summer correlation between cloud and coastline becomes disrupted, probably because the sea breeze cannot become established.

Double-headed arrows indicate bands that occur within a range of wind directions, and single-headed arrows where successive areas of convective activity follow similar paths.

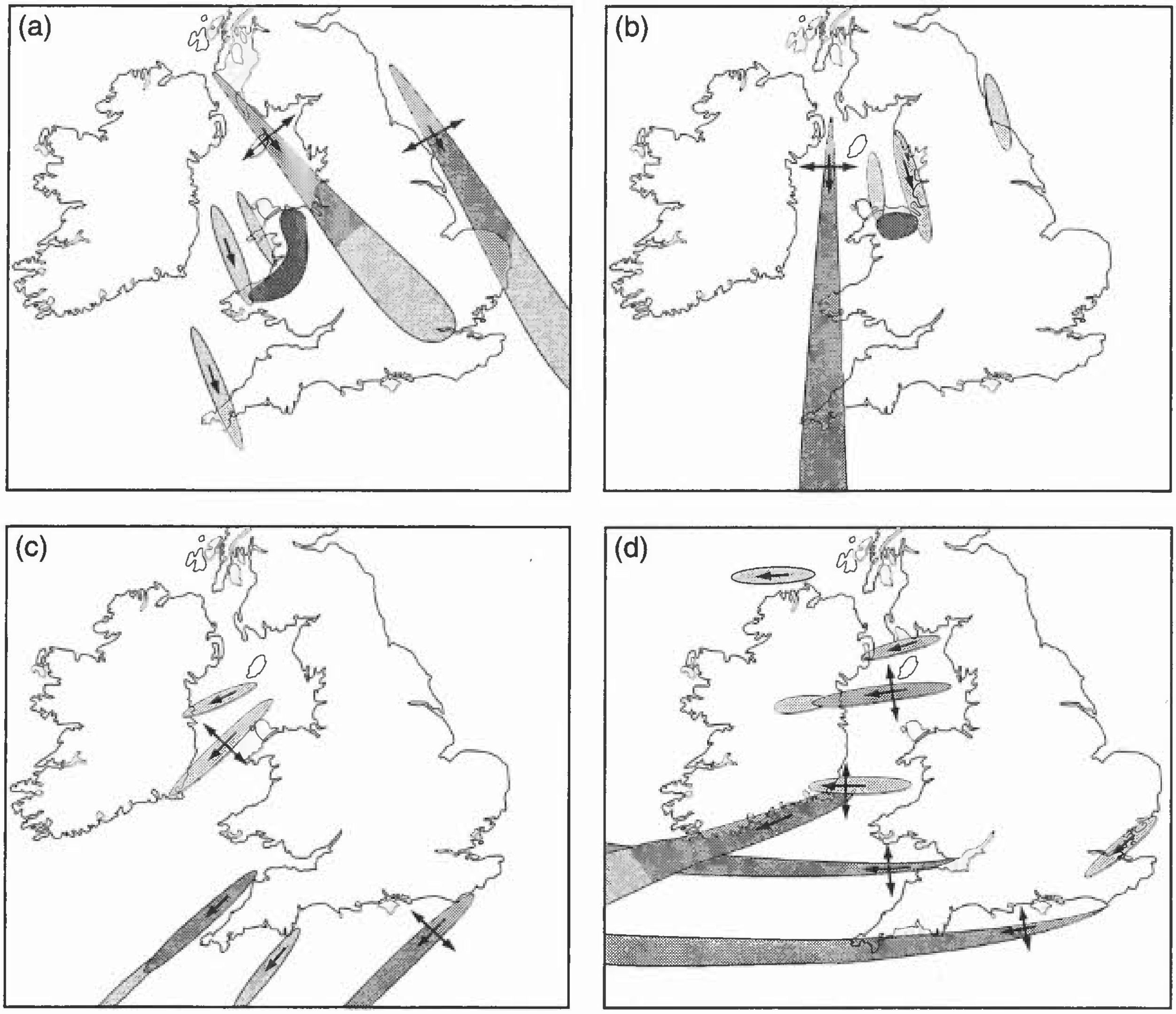

Figure 4.8. Winter convection generated due to air which is unstable to sea temperatures. (a) North-westerly airflow, (b) northerly airflow, (c) north-easterly airflow, and (d) easterly airflow.

4.7.2 Forecasting thunderstorms — instability indices

(a) *Boyden Index*

A measure of the instability below 700hPa is:

$I = (Z - 200) - T$

where Z = 1000–700 hPa thickness (dam); T = 700 hPa temperature (°C).

Thunder is probable if $I \geq 94/95$ (in the UK).

Forecasts should be made assuming that index isopleths move with the 700 hPa wind. A main advantage claimed for this method is its usefulness in mobile situations, regardless of whether fronts are involved. It should not be used in Mediterranean or tropical areas, or where the ground level is high.

Boyden (1966)

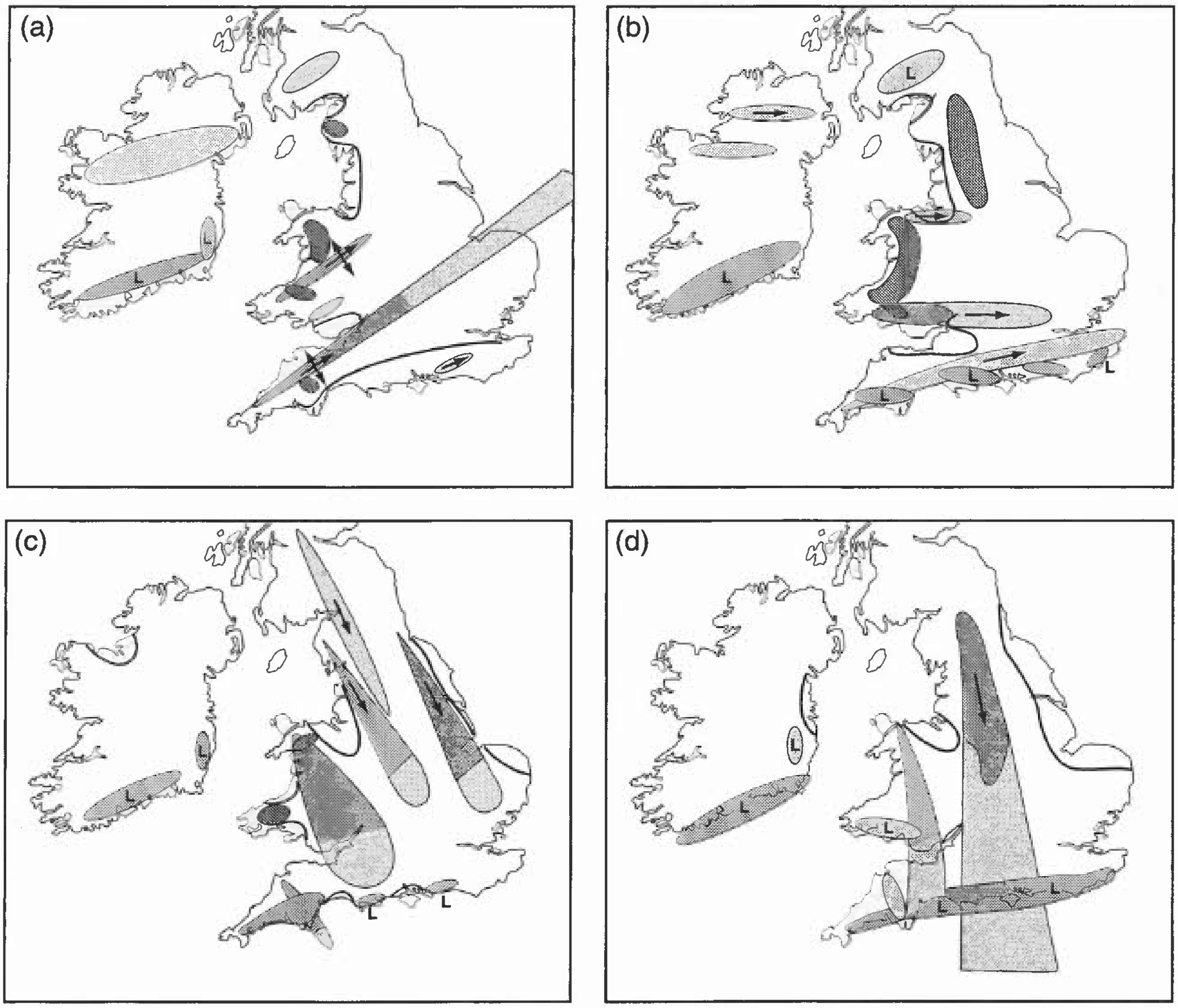

Figure 4.9. Summer convection generated due to air which is unstable to land temperatures. (a) South-westerly airflow, (b) westerly airflow, (c) north-westerly airflow, and (d) northerly airflow.

(b) *Rackliff Index*

The index is related to temperature as well as instability:

$$T = \theta_{w900} - T_{500}$$

where θ_{w900} is the 900 hPa wet-bulb potential temperature (°C); T_{500} is the 500 hPa temperature (°C).

In non-frontal situations significant showers accompanied by thunder are probable if $T \geq 29/30$ in the UK.

Rackliff (1962)

(c) *Modified Jefferson Index*

An amended form of Rackliff's index: values of that index for neutral stability air decrease almost linearly as WBPT increases. Jefferson's index gives an index independent of temperature (with same thunderstorm threshold value over a wide range of temperature). The formula was later amended to incorporate the 700 hPa dew-point depression to allow for middle-level humidity:

$$T_{mj} = 1.6\theta_{w900} - T_{500} - 0.5D_{700} - 8$$

where D_{700} = dew-point depression (°C) at 700 hPa.

Thunder is probable if T_{mj} = 27 to 28, in the United Kingdom although 26 to 27 is better in returning polar maritime air.

The formula can be reduced to $T_{mj} = \Delta T + X$ and computed more easily by using **Fig. 4.10,** where the value of $(-0.5D_{700} - 8)$ is replaced by the constant -11.

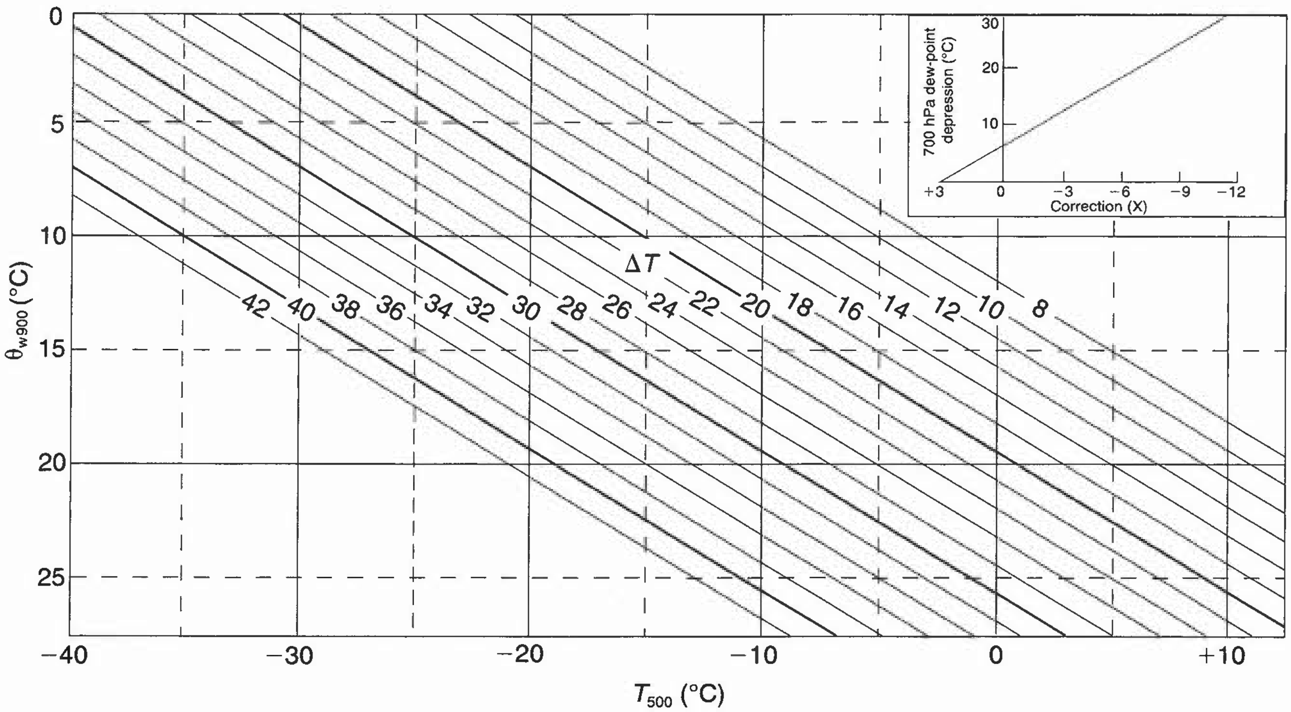

Figure 4.10. Computing the modified Jefferson Instability Index. See text for method of use.

(i) From the values of 500 hPa temperature (x-axis) and 900 hPa θ_w (y-axis), obtain a value of ΔT given by the sloping lines.

(ii) Correct this value of ΔT, according to the actual 700 hPa dew-point depression, using the small graph (top right) to obtain the correction X.

Jefferson (1963)

(d) *Potential instability index*

$$P = \theta_{w500} - \theta_{w850}$$

Thunder is possible if: P is ≤ -2 °C (summer)
P is $\leq +3$ °C (winter)

Bradbury (1977)

(e) *K index*

$\mathrm{K} = (T_{850} - T_{500}) + T_{d850} - (T_{700} - T_{d700})$, where: T_{d850} is 850 hPa dew-point temperature etc.
Thunder possible for $\mathrm{K} \geq 20$

George (1960)

4.7.2.1 Tests of different instability indices

Operationally indices offer only a guide to the degree of static instability at the time of a sounding. Other measures of convective development, such as radar and satellite imagery, as well as numerical weather prediction models, must provide supporting information.

Bradbury (1977)
Collier & Lilley (1994)

4.7.3 Hail

Deep and vigorous convection is required. The following criteria are a guide:

(i) Cumulonimbus tops are colder than −20 °C.

(ii) The 'parcel' path curve is warmer than the environment curve by 4 °C at some level (CV in **Fig. 4.1**) and gives cloud tops of 15,000 ft (4500 m) or more.

A similar method, based on the parcel curve, is:

(i) At the point where the curve reaches −20 °C measure the difference between this temperature and the environment temperature.

(ii) If this difference is:

	≥5 °C	forecast hail;
	from 5 °C to 2.5 °C	forecast soft hail or rain;
	if difference ≤2.5 °C	forecast rain.

(large hail requires a 'steady state', but not necessarily slow-moving, storm).

In both methods vertical wind shear exists between the base and top of cumulonimbus.

Browning (1963) **Ludlam (1980)**

4.7.4 Lightning

Lightning occurs in vigorous convective cloud; ice particles and hail are considered to play a key role in charge generation and, indeed, the vast majority of lightning emanates from thunderstorms extending well above the freezing level. However, there are well documented UK observations of lightning discharges from all-water clouds.

4.7.4.1 Detection and forecasting

(i) Often it is a reliable assurance that lightning is <u>not</u> going to occur that is required.

(ii) The Arrival Time Difference (ATD) system can detect both ground and cloud-to-cloud lightning strokes.

(iii) The current system detects about one third of all strokes, being limited by computer processing speed and station locations; the low false-alarm rate allows the detection of smaller thunderstorms earlier.

Atkinson et al. (1989) **Ludlam (1980)**
Lee (1986) **Mason (1971)**

4.7.4.2 Static risk for towed targets

Many incidents of aerial-towed targets glowing and then disintegrating are due, apparently, to static discharge associated with the following conditions:

(i) preceding a fairly lengthy dry spell;
(ii) too dry below cloud (Cu or Ac cast) level for thunder forecast;
(iii) cloud structure exhibits convective instability, but of shallow extent;
(iv) cloud form suggests a history of significant vertical development;
(v) aircraft is usually operating below cloud which has supercooled water or ice crystals present near the base;
(vi) some weak precipitation is often observed, usually slight showers or virga.

Both conducting and non-conducting tow lines have been used at the Aberporth Test and Evaluation ranges, but the influence of the conducting nature of the tow line has yet to be determined.

Static risk is also reported for helicopters but associated weather criteria await evaluation.

Aberporth Met. Office (1993) **Rogers (1967)**

4.7.5 Gust fronts

(i) Air from well-developed downdraughts spread out as a cold density current; leading edge convergence and uplift in this 'gust front' are marked by wind vector changes of, commonly, 10 m s^{-1} over a few hundred metres, with temperature falls of several degrees and pressure rises of up to 4 hPa. Downdraught temperature can be estimated from Fawbush & Miller, **Fig. 6.4**.

(ii) On average the gust front extends to 5 km ahead of the precipitation area, and so is often the precursor of rain. It is a region of potential daughter cell development (4.7.7), and may possibly be identifiable in imagery by an arc-cloud formation.

Bader et. al (1995), Chapter 6.4.3

4.7.6 Squall lines

Associated with *thunderstorms* as follows:

(i) Lines of Cb may form into 'squall lines', typically 30 km wide and 200 km long.

(ii) As they advance, moving in the direction of the shear, the systems force high θ_w surface air to the level of free convection, where it joins the updraught, cooler air from aloft being brought down behind the system.

Associated with *cold active fronts* with rearward-sloping warm conveyor belt (7.1):

(i) Line convection, marked wind shift and sharp pressure changes indicate the likelihood of a squall.

(ii) The severity of the squall probably depends on:

Wind speed in the low-level jet ahead of the cold front.
Wind speeds in the medium-level layers near the frontal surface.

See also 6.2.2.3 — *Gust forecasting in strong wind situations.*

HWF (1975) Chapter 16
Ludlam (1980)

4.7.7 Single- and multi-cell development

4.7.7.1 Effect of vertical wind shear

(i) With *no shear,* the updraught and downdraught in a cumulonimbus are coincident. The precipitation falls through the updraught which is weakened by the drag of the falling precipitation, and the cloud quickly decays.

(ii) With *wind speeds varying with height but no directional shear,* the updraught is tilted and the precipitation falls down beside the updraught. The cloud persists longer than in a no-shear situation. Eventually the updraught may be cut off at the surface by the spreading out of the cold downdraught beneath the storm.

(iii) With *directional and speed shear*, some complex mesoscale storm systems may result. These may develop into self-generating steady-state systems, multi-cell storms, which can persist for many hours, quite independently of any surface heating.

4.7.7.2 Changing wind direction and speed with height

(i) Under conditions of directional and speed shear the storm-cloud motion vector is found within the triangle formed by the wind vectors LMH (**Fig. 4.11(a)**); the downdraught falls to one side of low-level inflow (**Fig. 4.11(b)**).

(ii) Successive daughter cells may be generated where the surface outflow from the downdraught meets the inflow. Very high precipitation totals may result if cells are continuously generated in the same position. The storm area moves to the right of the track of individual cells if the wind veers with height, the most common case (**Fig. 4.12**).

(iii) The arrival of this mesoscale convective complex (MCC) system is preceded by extensive anvils, visible on satellite imagery as vast, long-lived areas which hide the elongated squall line producing them.

4.7.7.3 MCC systems: characteristics

Whether the MCC is a 'forward' or 'backward' propagating system will depend on whether new developments occur downwind or upwind (respectively) of the existing MCC system relative to the mean tropospheric wind.

Table 4.6. Characteristic properties of MCC systems near the British Isles

Dimensions	overall ≈300 km, with embedded cells ≈50 km
Cloud-top temperature	<-30 °C over wide area, locally <-50 °C
Duration as an active system	≈12 hours

Bader et al. (1995) Chapter 6

4.7.7.4 Supercells

(a) A very small proportion of MCC systems will develop into 'supercells' that propagate continuously with highly organized internal circulation with low θ_w downdraught coexisting with the high θ_w updraught, forming a mini-frontal system.

(b) A distinguishing feature is that cell movement is in a different direction from winds at any particular level (i.e. the storm motion vector falls outside the triangle formed by the wind vectors LMH, **Fig. 4.11(a)**, possibly due to effects induced by the rotating of the cell.

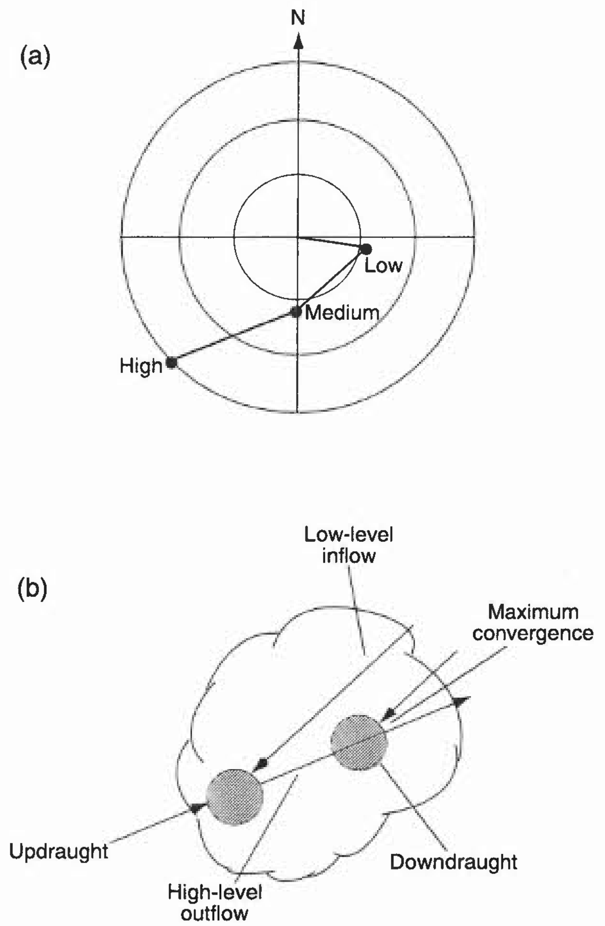

Figure 4.11. (a) Wind environment for multi-cell storm represented on a hodograph, and (b) configuration of updraughts and downdraughts from consideration of system-relative winds.

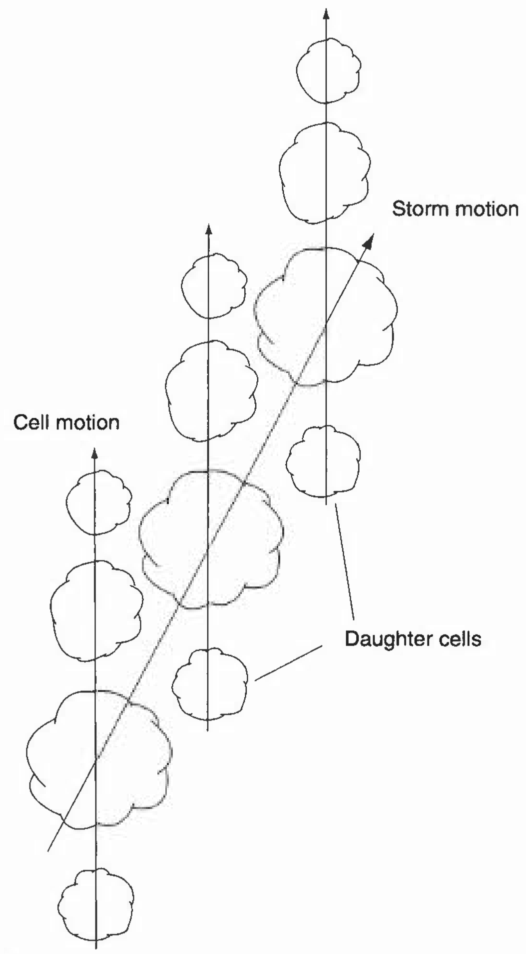

Figure 4.12. Multi-cell storm travelling to right of individual cells.

(c) The 'supercell' is responsible for the most severe summertime thunderstorms with localized flooding. Southern England is a favoured location with a low-level flow of warm, continental south-easterly winds overridden by cooler, oceanic south-westerlies aloft. In extreme cases with light winds storms may persist at a particular place for several hours with heavy rain and hail, although such storms are not common in the UK.

(d) Typical synoptic-scale environment for supercell storms:
   (i) Strong instability with parcel theory indicating >4 °C excess buoyancy at 500 hPa.
   (ii) Strong mean sub-cloud winds (order of 10 m s^{-1}).
   (iii) Strong environmental shear through cloud layer of 2.5 to 4.5 m s^{-1} per km (hence strong upper winds).
   (iv) Strong veer of wind with height.

Bader et al. (1995), Chapter 6
Browning & Ludlam (1962)
Ludlam (1980)

4.7.7.5 CAPE and development of severe storms

Severe storms may spawn 'daughter cells' to their north-east flank, due to convergence; the maximum area of convective activity (as tracked, for example, by radar) will move to the right of the individual cell motion, typically by 20 to 30° **(Fig. 4.12)**.

Development of severe storms depend on:
(i) Vertical wind shear through the depth of the convective layer ΔU **(Figs 4.11 and 4.13)**.
(ii) Large quantities of Convectively Available Potential Energy (CAPE), which can be released through convection. On a tephigram this energy is represented by the area between the environment curve and the path curve of a

rising parcel (ATUVWCB in **Fig. 4.1**). The severity of the storms depends on whether the CAPE is released by many small cumulonimbus clouds or by a few giant ones.

(iii) A measure of whether or not a storm will be single- or multi-cell is estimated from:
$R > 3$ — storms likely to be multi-celled;
$0.5 < R < 1$ — single cell likely,
where: R (the bulk Richardson number) $=$ CAPE/$[0.5(\Delta U)^2]$, i.e a measure of the wind shear required to organize flow, against buoyancy forces tending to disrupt the flow.

Collier & Lilley (1994)
Galvin et al. (1995)
Ludlam (1980)

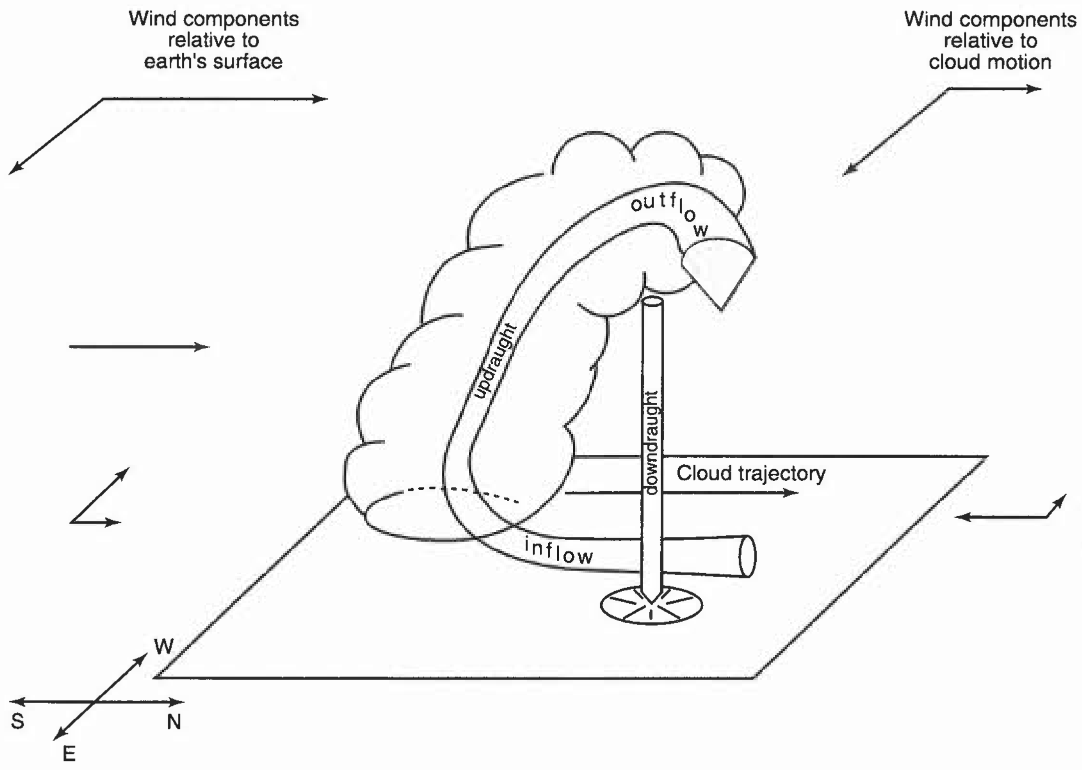

Figure 4.13. Cumulonimbus cloud in directional shear.

4.7.8 Forecasting thunderstorms — synoptic features

The objective forecasting techniques should not be used in isolation. Their value is greatly enhanced when they are used in conjunction with features which can be analysed on synoptic charts. Upper-air soundings alone do not always reveal the extent of the thunder risk, the soundings available before a thundery outbreak failing to show exceptional instability.

Consideration should be given to factors likely to release the energy available for convection, such as:

(i) The position and movement of upper-level troughs or lows. Thunderstorms are much more likely to break out along or ahead of an upper trough, or near an upper low.
(ii) The existence and movement of low-level convergence lines, such as surface fronts, or sea-breeze fronts.
(iii) Areas of high ground which may become very warm by day in the summer months.

Other useful synoptic tools are:

(iv) Dew-point analysis: the movement of a tongue of air with high dew points may help to define a thunder-threatened area.
(v) Charts of θ_w at 850 hPa: these serve a similar purpose as surface dew-point analyses.
(vi) Charts of the difference between θ_w at 500 and 850 hPa: these show where θ_w decreases with height over a significant depth of the atmosphere, and potential instability exists. Be on the lookout for $\theta_w \geq 18$ °C.
(vii) Cyclonic curvature of surface isobars: minor troughs are associated with convergence and enhanced convection.

4.7.8.1 Conditions favouring severe thunderstorms

This section is concerned only with the synoptic-scale setting in which the storms occur. Over Britain, severe storms are often associated with a cold front over, or close to, north-western parts with an attendant upper trough to the west. Under these conditions:

(i) Baroclinicity ensures a fairly strong south-west flow at upper levels.
(ii) At medium levels, air of Saharan origin passes over the Spanish plateau.
(iii) This air limits convective depth over south-west France and θ_w values attain as much as 24 °C there, leading to a heat low over France.
(iv) Overnight the low-level high θ_w air is advected northwards around the low, approaching southern England from the south or south-east next day.
(v) The warm, dry air of low θ_w (the 'Spanish Plume') has undergone mass ascent with its northward progression. It cools, moistens, sometimes producing bands of Ac castellanus and thundery outbreaks as it approaches the UK from across the Bay of Biscay.
(vi) By the time it reaches the UK the 'lid' produced by warming over the Spanish plateau has been removed; potential instability can then be realised.

A representation of synoptic-scale air flow associated with outbreaks of severe convection over southern Britain is shown in **Fig. 4.14**. Sometimes thunderstorms break out over northern France and advect northwards, becoming more organized and severe on encountering the stronger upper-level winds. Thunderstorms are most likely to break out within the tongue of highest θ_w air and where there is low-level convergence due to isobaric troughing or mesoscale effects such as sea-breezes.

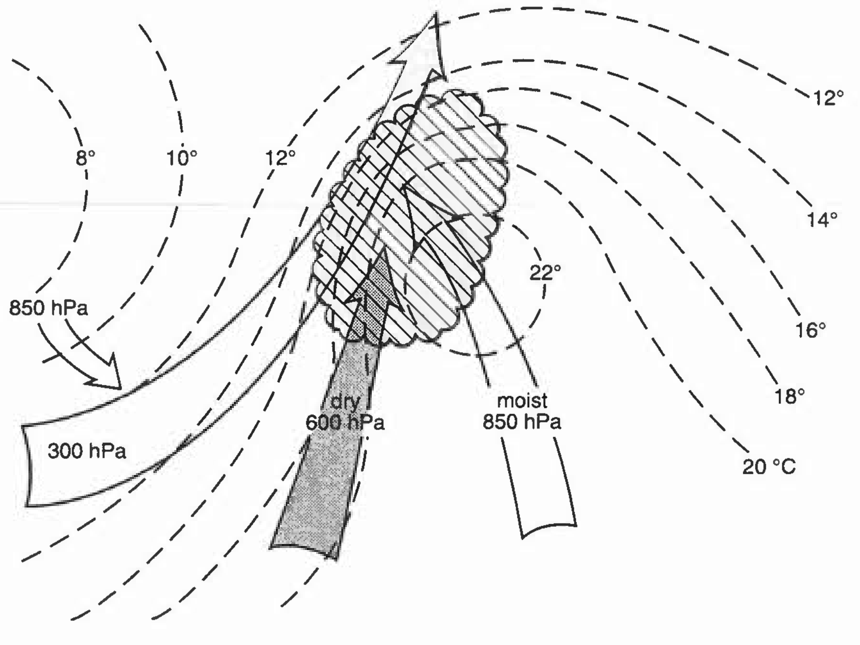

Figure 4.14. A schematic diagram of the three main synoptic-scale currents associated with the development of severe thunderstorms over southern England, together with typical isopleths of θ_w at 850 hPa. The hatched area indicates the the location of the storms.

Pointers to forecasting thunderstorms over southern England:

(i) *Upper-air soundings*
Some of the most severe and widespread outbreaks of thunder have occurred some 12–18 hours after a sounding showed a layer of warm air capping a shallow convective layer. This cap can prevent the early release of the convective energy, allowing it to build up with continued surface heating and be released explosively in one big storm later.

(ii) *Advection of low-level moisture*
Surface dew points should exceed 13 °C, but often reach 18 °C or more in severe storms over southern England. At 850 hPa, similar values of θ_w are experienced. The winds at this level are usually in the sector SE–SSW, with a speed of 20–30 kn, often in a narrow tongue.

(iii) *Medium-level advection of dry air*
With potential instability present, θ_w values at 500 hPa may be 2–5 °C lower than at 850 hPa. Winds at 500 hPa should be 20–40° veered from those at 850 hPa, with speeds of 35–50 kn.

(iv) *Upper-level strong winds*
Further veering above 500 hPa, with 300 hPa winds in the sector SSW–W and speeds 50–85 kn, are good conditions for positioning the downdraughts in the favoured position for generating severe storms.

(v) *Positive vorticity advection (PVA) at 500 and 300 hPa*
This usually occurs in the region in advance of an upper trough or upper low. Isopleths of vorticity cross the contour lines at an angle of 30° or more.

(vi) *'Dry lines'*
Mesoscale areas of dry air within an otherwise moist air mass, and the resulting production of strong horizontal moisture gradients (*dry lines*) by deformation in the air ahead of a (summer) cold front, may provide a preferred position for triggering intense thunderstorm activity in the already unstable air. Satellite, radar and model data should distinguish the deep, dry areas well before the outbreak.

Bader et al. (1995), Chapter 6
Grant (1995)
Ludlam (1980)
Morris (1986)
Scorer & Verkaik (1989)
Young (1995)

BIBLIOGRAPHY

CHAPTER 4 — CONVECTION AND SHOWERS

Aberporth Met. Office, 1993: Local Staff Instructions.

Atkinson,N.C, Blackburn, M.R. and Kitchen, M., 1989: Wide area lightning location using the UK Met. Office arrival time difference system. International Conference on Lightning and Static Electricity, University of Bath, 26/28 September 1989.

Bader, M.J., Forbes, G.S., Grant, J.R., Lilley, R.B.E. and Waters, J., 1995: Images in weather forecasting. Cambridge University Press.

Booth, B.J., 1978: On forecasting dry thermals for gliding. *Meteorol Mag*, **107**, 48–61.

Booth, B., 1980: Unusual wave flow over the Midlands. *Meteorol Mag*, **109**, 313–324.

Boyden, C.J., 1966: A simple Instability Index for use as a synoptic parameter. *Meteorol Mag*, **92**, 198–210.

Bradbury, T.A.M., 1977: The use of wet-bulb potential temperature charts. *Meteorol Mag*, **106**, 233–251.

Bradbury, T.A.M., 1978: Weather conditions for long glider flights over England. *Meteorol Mag*, **107**, 340–353.

Bradbury, T.A.M., 1990: Links between convection and waves. *Meteorol Mag*, **119**, 112–120.

Bradbury, T.A.M., 1991a: Thermal prediction from the tephigram. *Sailplane & Gliding*, **June/July**, 122–126.

Bradbury, T.A.M., 1991b: Meteorology and flight. A & C Black.

Browning, K.A., 1963: The growth of large hail within a steady updraught.*QJR Meteorol Soc*, **89**, 490–506.

Browning, K.A. and Ludlam, F.H., 1962: Airflow in convective storms. *QJR Meteorol Soc*, **88**, 117–135.

Browning, K.A., Eccleston, A.J. and Monk, G.A., 1985: The use of satellite and radar imagery to identify persistent shower bands downwind of the North Channel. *Meteorol Mag*, **114,** 325–331.

Collier, C.G. and Lilley R.B.E., 1994: Forecasting thunderstorm initiation in NW Europe using thermodynamic indices, satellite and radar data. *Meteorol Appl,* **1**, 75–84.

Galvin, J.F.P., Bennett, P.H. and Couchman, P.B., 1995: Two thunderstorms in summer 1994 at Birmingham. *Weather*, **50,** 239–250.

George, J.J, 1960: Weather forecasting for Aeronautics. Academic Press (K Index).

Grant, K., 1995: The British 'dry line' and its role in the genesis of severe local storms. *J Meteorol*, **20**, 241–259.

Hill, F.F., 1983: The use of average annual rainfall to derive estimates of orographic enhancement of frontal rainfall over England and Wales for different wind directions. *J Climatol*, **3**, 113–129

Jefferson, G.J., 1963: A further development of the Instability Index. *Meteorol Mag*, **92**, 313–316.

Lee, A.C.L., 1986: An operational system for the remote location of lightning flashes using a VLF arrival time difference technique. *J Atmos and Oceanic Technol*, **3**, 630–642.

Local Weather Manual for Southern England, 1994: Meteorological Office.

Ludlam, F.H., 1980: Clouds and storms. Pennsylvania State University Press.

MacIntosh, D.H. and Thom, A.S., 1972: Essentials of meteorology. London, Wykeham Pubs.

Mason, B.J, 1971: The physics of clouds (2nd edition). Clarendon Press, Oxford.

Met O 6, 1989: Notes on the preparation of gliding forecasts for RAF gliding schools. Meteorological Office (Met O 6), Unpublished.

Meteorological Glossary, 1991: London, HMSO.

Monk, G.A., 1987: Topographically-related convection over the British Isles. *In* Workshop on 'Satellite and Radar Imagery Interpretation' at the Meteorological Office College, Shinfield, Ed: M. Bader and A. Waters; pub: EUMETSAT.

Morris, R.M., 1986: The Spanish plume — testing the forecaster's nerve. *Meteorol Mag*, **115**, 349–357

Observer's Handbook, 1982: London, HMSO.

Orographic Processes in Meteorology (Pre-prints), 1993: Summer School, Meteorological Office College.

Pettersen, S., Knighting, E., Jones, R.W. and Herlofson, N., 1945: Convection in theory and practice. *Geofys Publ*, Oslo, **16**, No. 10.

Rackliff, P.G., 1962: Applications of an Instability Index to regional forecasting. *Meteorol Mag*, **91**, 113–120.

Rogers, M.E., 1967: Interim report reviewing the present position on helicopter static electrification. RAE Technical Report 67292.

Scorer, R.S. and Verkaik, A., 1989: Spacious skies. David and Charles.

Young, M.V., 1995: Severe thunderstorms over south-east England on 24 June 1994: A forecasting perspective. *Weather*, **50,** 250–256.