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Stratocumulus: an introductory account

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Summary

This article on stratocumulus is one of a series of teaching papers on mesoscale meteorology developed at the Meteorological Office College. It describes the important physical and dynamical aspects of stratocumulus formation and dissipation, highlighting the roles of radiation, turbulence, subsidence and cloud microphysics. The aim is to provide simple conceptual models which will help meteorologists develop an understanding of this type of cloud.

1. Introduction

Stratocumulus is very common around the United Kingdom and usually occurs in large sheets, sometimes covering areas of about 10^6 km^2. Fig. 1 shows how often stratocumulus of some sort is present in the sky.

The presence of low-level layer cloud significantly affects the radiation balance of the lower atmosphere, thereby modifying both the structure of the boundary layer and the surface energy balance. Consequently, forecasts of boundary-layer phenomena, such as fog formation and dispersal, maximum and minimum temperatures, surface conditions, etc. are highly sensitive to the presence of stratocumulus.

It is difficult to forecast accurately its disperal and possible re-formation after a temporary clearance. In fact relatively little progress has been made towards developing reliable forecasting techniques, making it all the more important that forecasters should understand the physical and dynamical processes occurring in sheets of stratocumulus.

The object of this paper is to provide this physical insight without obscuring basic understanding with too much detail.

2. General features

Most data on the structure of stratocumulus have been obtained in the mid-latitudes from single cloud layers (e.g. stratocumulus near high pressure regions). Subtropical and arctic stratocumulus, as

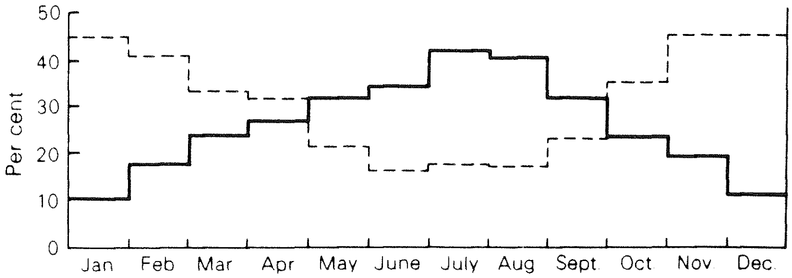

Figure 1. Frequency of occurrence of stratocumulus (pecked line) and stratocumulus mixed with cumulus (bold line) in daytime observations for Boscombe Down 1960–69.

well as multiple layers of stratocumulus, have been less well studied so some caution must be exercised in applying the present results to such cases.

Stratocumulus formation is usually associated with either (a) the cooling or moistening (or both) of the boundary layer, or (b) the spreading out of cumulus beneath an inversion. The general features are best illustrated by reference to an example. Fig. 2(a) shows vertical structure measured by the Meteorological Research Flight Hercules aircraft during a slow descent (solid lines) together with a simplified representation (dotted lines). There are five regions of interest, namely: just above cloud (region A), cloud top (region B), within cloud (region C), cloud base (region D) and below cloud (region E).

Region A – just above cloud

In this region the air is warm and dry, often through subsidence. The air is stable and there is little or no turbulence.

Region B – cloud top

At the top of the cloud layer there is a marked inversion and hydrolapse. This is particularly well defined by the temperature profile. The boundary between the cloud layer and the subsiding air (which is several degrees warmer) has been observed to be as little as a few metres thick, but more generally it is a few tens of metres. The consequent large temperature, and hence density, gradient makes the atmosphere very stable and local perturbations are strongly damped, so that the cloud top is usually fairly flat, although small-amplitude gravity waves are often present.

Wind shear is usually confined to the inversion layer at the cloud top, where it may be large.

Region C – within cloud

Within cloud the air is well mixed, as shown by the temperature profile in Fig. 2(b), (and hence there is little wind shear) and has a constant wet-bulb potential temperature. Since in this example there is negligible precipitation, the total water content of a parcel of air rising through the cloud is conserved.

In discussing the water content of the atmosphere there are two important quantities: the amount held as vapour and the amount held as cloud water. The relation between them is best seen by an example. Consider a parcel of air at the surface at 10 °C with 6 g kg^{-1} of water vapour. Lift this parcel of air, without entrainment, without loss of water through precipitation and without any exchange of heat to or from the environment, from 1000 mb to 800 mb. The cloud base is at 945 mb.

As the parcel rises through the atmosphere above the cloud base the amount of cloud water increases (roughly linearly) while the amount of vapour decreases. In fact the amount of cloud water increases at a value of about 0.7 g kg^{-1} km^{-1} or 1.0 g m^{-3} km^{-1} (the density of air at 1000 mb is about 1.2 kg m^{-3}). This value is referred to as the adiabatic liquid water content. In practice the liquid water content of stratocumulus is always slightly below this value owing to the entrainment of dry air at cloud top (this is

Height (mb)	Humidity mixing ratio of air ($g\ kg^{-1}$)	Cloud liquid water content ($g\ kg^{-1}$)	Total ($g\ kg^{-1}$)
800	3.9	2.1	6.0
850	4.6	1.4	6.0
900	5.3	0.7	6.0
950	6.0	0.0	6.0
1000	6.0	0.0	6.0

revealed in Fig. 2(a) by the departure from the adiabatic value, shown by the dotted line, in the upper part of the cloud). Precipitation would also lower the liquid water content.

Icing on aircraft can be a major hazard in stratocumulus. This will occur with subzero temperatures (below about 10 °C if carburettor icing is considered) and high liquid water content. The latter reaches its maximum value near the cloud top. When temperatures fall below −20 °C most of the water is in the form of ice particles which bounce off the aircraft, i.e. there is little danger of icing. The main danger is in the temperature range 0 to −15 °C when there are many supercooled water droplets which freeze on impact with an airframe.

Region D – cloud base

The transition to clear air at the base of the cloud layer is poorly defined. In contrast to cloud top, the cloud base region has no strong temperature gradients and stability to vertical motion is weak. In consequence, the cloud base is diffuse and large perturbations can grow, as is evident from the rolls that are frequently observed at the base of stratocumulus sheets.

Region E – below cloud

Below cloud the air is also well mixed, with the same wet-bulb potential temperature as the cloud layer. In fact on many occasions the within-cloud and below-cloud regions may be treated as one, even though the potential temperature profile shows a discontinuity at the cloud base.

3. What controls the development of stratocumulus?

Stratocumulus results from interactions between processes having widely differing length and time-scales, in particular:

(a) Synoptic-scale subsidence maintains the inversion and tends to lower the cloud top.
(b) The stratocumulus lifetime of several days implies that radiative effects are important.
(c) Turbulent mixing raises the cloud top. ((a) and (c) are often in near balance.)
(d) The detailed structure of the cloud is defined by the microphysical processes.

We shall consider these four processes separately, then discuss their interaction.

(a) *Motion at the cloud top*

The cloud top is not an impenetrable surface — it merely marks the boundary between clear air and air containing drops. When cold air sinks away from the cloud top some air is drawn down from above the cloud and mixes with the cloudy air. In particular, the downdraughts formed at cloud top must entrain some warm dry air from above the inversion some of which would be replaced by air from within the cloud. It is difficult to observe this process (but see Fig. 2(a), humidity mixing ratio) as it takes place

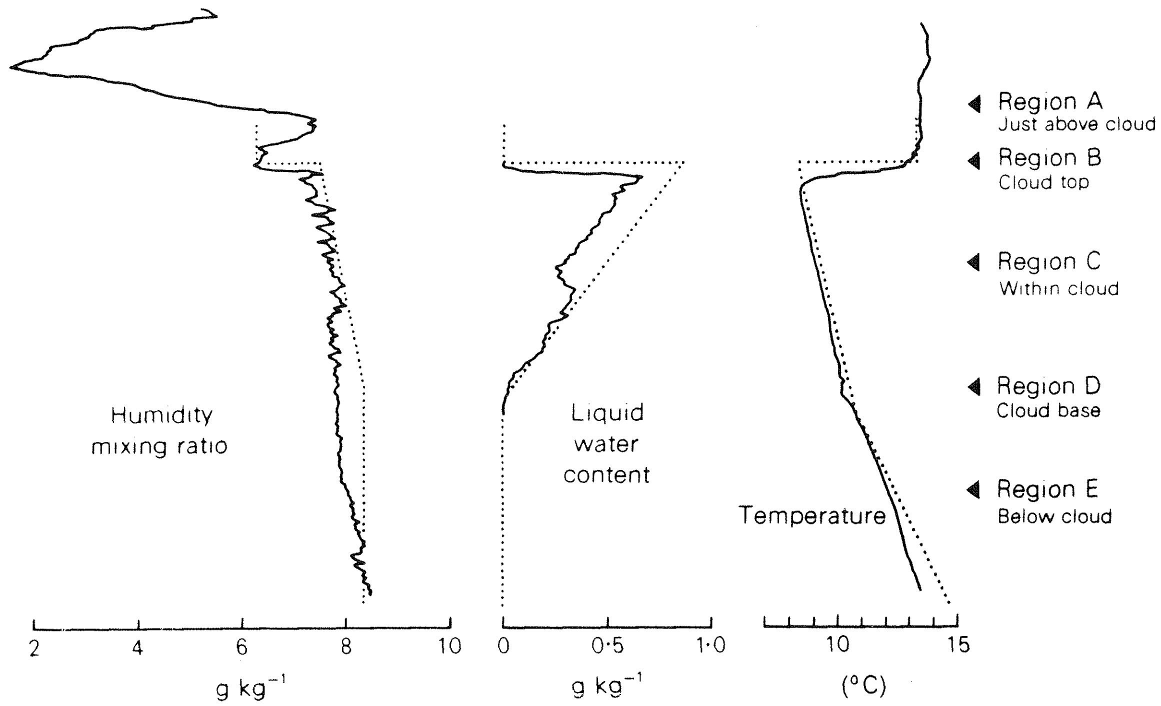

Figure 2(a). An example of the humidity mixing ratio, liquid water content and temperature profiles in stratocumulus. The solid lines show the structure as measured by the Meteorological Research Flight C130 aircraft. The dotted lines show an idealized representation for stratocumulus.

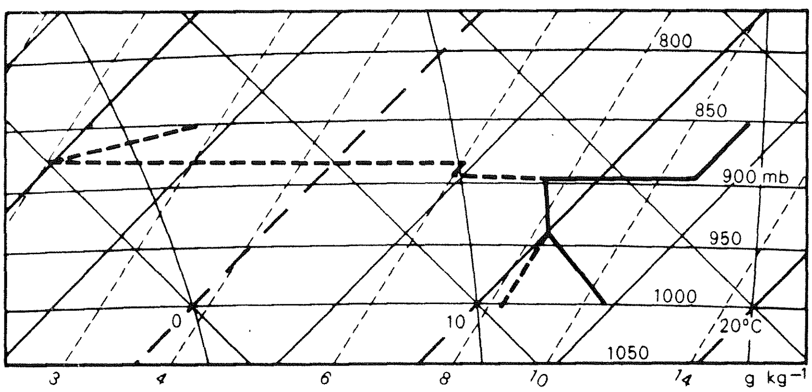

Figure 2(b). Plotted ascent corresponding to the idealized profiles in Fig. 2(a).

within a few tens of metres of cloud top, but over a period of time the effect is to deepen the mixed layer and the cloud top rises. Initially, the entrainment of dry air into the cloud top may decrease the liquid water content of the cloud while the injection of cloudy air into the above-cloud region will produce little visible difference. However, on a longer time-scale the process moistens the region just above cloud top and eventually, given a steady supply of moisture from the ground, the region will reach saturation and the cloud thicken. This effect is opposed by the large-scale anticyclonic subsidence, though the two need not be in balance. In consequence, the cloud layer will locally rise and fall, depending on whether one is stronger than the other. Aircraft observations show that frequently cloud top (and cloud base as well) slopes by 50–100 m per 100 km.

(b) *Radiative effects*

In general, radiative transfer takes place at all wavelengths but in the atmosphere we are primarily concerned with long-wave radiation (4–40 μm) and short-wave radiation from the sun (0.3–3 μm).

(i) Long-wave radiation

Once the cloud is sufficiently deep, typically about 150 m around the United Kingdom, the cloud layer acts approximately like a black body. The main long-wave properties can then be described by considering the radiative balance at cloud top, cloud base and within the cloud layer.

At cloud top, the cloud radiates (loses) energy according to its temperature (about 275 K) but receives very little from the overlying 'transparent' layers of the atmosphere. Cooling rates are therefore large, about 5–10 K per hour, and occur within the top few tens of metres of the cloud. Within the cloud there is little net radiative transfer as all the cloud droplets have a similar temperature, thus energy radiated away from a given volume inside the cloud is almost exactly balanced by that received from surrounding regions. At cloud base, since the ground temperature is usually a few degrees higher than cloud temperature, there is slight warming. Long-wave radiative transfer continues throughout the day and night, and its effect is summarized in Fig. 3(a).

(ii) Solar radiation

It is more difficult to summarize the role of solar radiation because of its variation with time of day, latitude, season, etc. In addition, the albedo of the cloud layer (i.e. the percentage of the incident flux reflected back to space) varies from 40–90%, depending on cloud thickness, drop size, solar elevation, etc. Of that which penetrates, some is absorbed by the cloud (13–15%) while the remainder reaches the ground. Another feature of the solar radiation is that, since it has a shorter wavelength, the depth over which it is absorbed by the cloud is much greater than that for long-wave radiation. Fig. 3(b)

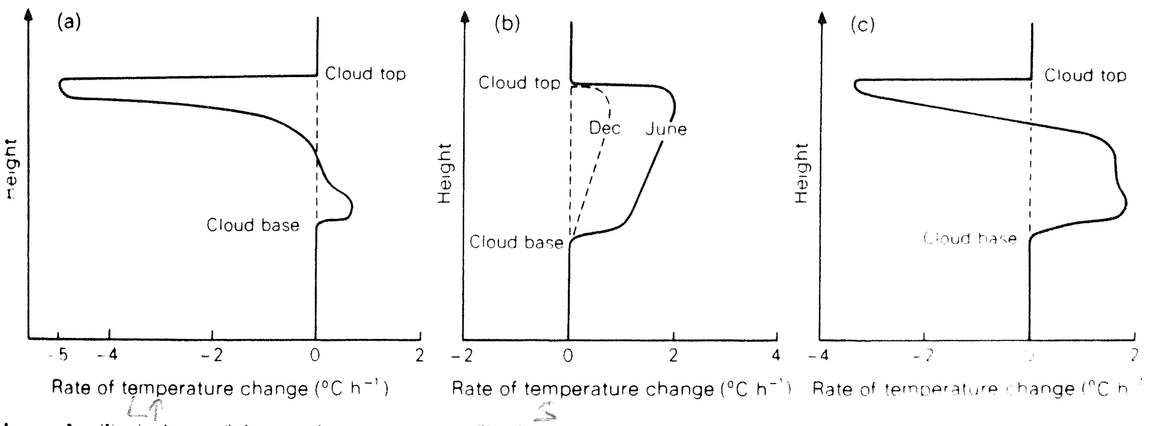

Figure 3. Typical rate of change of temperature profiles for stratocumulus showing (a) the net effect of long-wave radiation, (b) the net effect of solar radiation in the United Kingdom in June (solid line) and December (dashed line) and (c) the combined effect of long-wave and solar radiation in the United Kingdom in June.

shows the net warming of the cloud in June at midday in the United Kingdom (latitude 52 °N). In winter this would be reduced by a factor of about four owing to the low elevation of the sun (dashed line in Fig. 3(b)) and at night would vanish. The combined effect of long-wave radiation and insolation (for the United Kingdom in June) is shown in Fig. 3(c). Note that in June at midday in the United Kingdom the amounts of solar radiation absorbed and long-wave radiation emitted by the cloud are nearly equal, although their spatial distribution is different. In consequence, stratocumulus may be expected to display significant diurnal changes, especially during the summer months.

(c) *Turbulence and entrainment*

The distribution of radiative heating and cooling, either by day or night, would, by itself, generate convective instability. This is released through turbulent motions which are observed to take the form of cold downdraughts descending from the vicinity of cloud top, with associated warm compensating updraughts. Thus cold air is brought down further into the cloud from the cloud top while warm air ascends. Since the rate of transfer of heat depends on the temperature difference within the cloud the system is self-regulating and, unless the cloud is dissipating or forming, effectively spreads the temperature changes due to radiation throughout the cloud.

However, as the long-wave cooling and solar heating will rarely be exactly in balance, in general the cloud experiences a net cooling. This will be discussed later, in more detail.

(d) *Microphysical structure*

As already stated, the liquid water content within stratocumulus increases roughly linearly with height at slightly below the adiabatic value. However, aircraft measurements show that there is usually little variation of droplet concentration with height (i.e. the number of cloud droplets stays the same but the drops become bigger towards the top of the cloud). The deeper the cloud layer, the larger the drops that form within it, and therefore there exists some critical depth at which drizzle (droplets more than about 100 μm) can form. It would be useful to derive a simple expression for this critical depth.

From microphysical theory there are two relevant facts:

(i) droplets less than about 20 μm radius grow only by condensation, and

(ii) droplets more than about 20 μm radius can grow by coalescence provided they are surrounded by droplets of different size.

Therefore, to produce drizzle, a few droplets must first grow, by condensation, to 20 μm. Thereafter growth by coalescence to produce drizzle is relatively rapid, although the process is complicated by the random nature of droplet collisions. Neighbouring droplets are not all the same size so it is helpful to define a mean droplet radius and to consider a size distribution about that radius. For example, if the mean radius is 10 μm then in stratocumulus one would expect to find drops ranging in size from about 5 to 15 μm, although these limits are not rigid. Thus a few 20 μm particles appear when the mean radius is about 15 μm. Therefore, knowing the critical mean radius (15 μm) at which 20 μm droplets first appear (i.e. the critical mean radius from which drizzle can subsequently form) and the liquid water content, the critical depth of cloud can be calculated provided that we know the number concentration of the cloud droplets.

For example, let the number of droplets be N cm^{-3}. Then the liquid water within a cloud is given by $N \times$ (mean mass of the drops). Taking the mean radius as 15 μm the liquid water content equals

$$N \frac{4\pi}{3}\left(\frac{15}{10^{6}}\right)^{3} \times 10^{12}\ \mathrm{g\ m^{-3}}. \qquad \ldots\ \ldots\ \ldots\ \ldots\ \ldots\ \ldots\ (1)$$

Within stratocumulus the adiabatic liquid water content varies with height at a rate of about

1.0 g m^{-3} km^{-1} (section 2) therefore at the top of a cloud of depth d metres the liquid water content is:

$$d\,10^{-3}\ \mathrm{g\ m^{-3}}. \qquad \ldots \quad (2)$$

Equating (1) and (2) we have (very roughly)

stel hoeveelheid druppelwater gelijk aan [illegible] ontstaat vloeistof

$$d = 10N\ \mathrm{m}.$$

This relationship is summarized in Table I. Aircraft measurements have shown that N ranges from about 50 for clean maritime air to greater than 500 for industrial areas in continental regions. There is no way of measuring N from synoptic data but a good estimate can be made by considering the source of an air mass and its subsequent trajectory.

Table I. *An estimate of the depth of stratocumulus containing water droplets only required to produce drizzle at cloud base in a layer with cloud-top temperature above about −5° C*

Air mass	Number of water droplets (N cm^{-3})	Minimum depth of cloud to produce drizzle at cloud base (m)
Very clean maritime	50	500
Maritime	100	1000
Continental	200	2000
Industrial continental	250	2500

Whether or not the drizzle reaches the ground will depend on the humidity of the air below cloud base and local orographic effects that result in increased vertical motion.

Table I refers to stratocumulus cloud containing only water droplets. It provides only a very rough guide since droplet coalescence is a statistical process which has been poorly represented in the above calculations. In real clouds some drops, by chance, grow faster than others; also the effects of turbulence can 'recycle' certain drops allowing them a larger than average time in which to grow. In reality, rainfall production is a very complex process which is difficult to represent simply.

If the stratocumulus layer has a cloud-top temperature below about −5 °C, ice processes can become important and greatly enhance the production of precipitation, thereby reducing the depths in Table I. However, whether or not ice is present, maritime stratocumulus is always more likely to produce precipitation than continental stratocumulus.

Fig. 4 provides a summary of the processes discussed so far as well as indicating energy input from the surface.

4. The interaction of different scales of motion

(a) *Nocturnal effects*

Consider a sheet of stratocumulus in steady state during the day, with a radiation balance as shown in Fig. 3(c). At dusk the solar radiation ceases and the structure of the heating/cooling changes to that

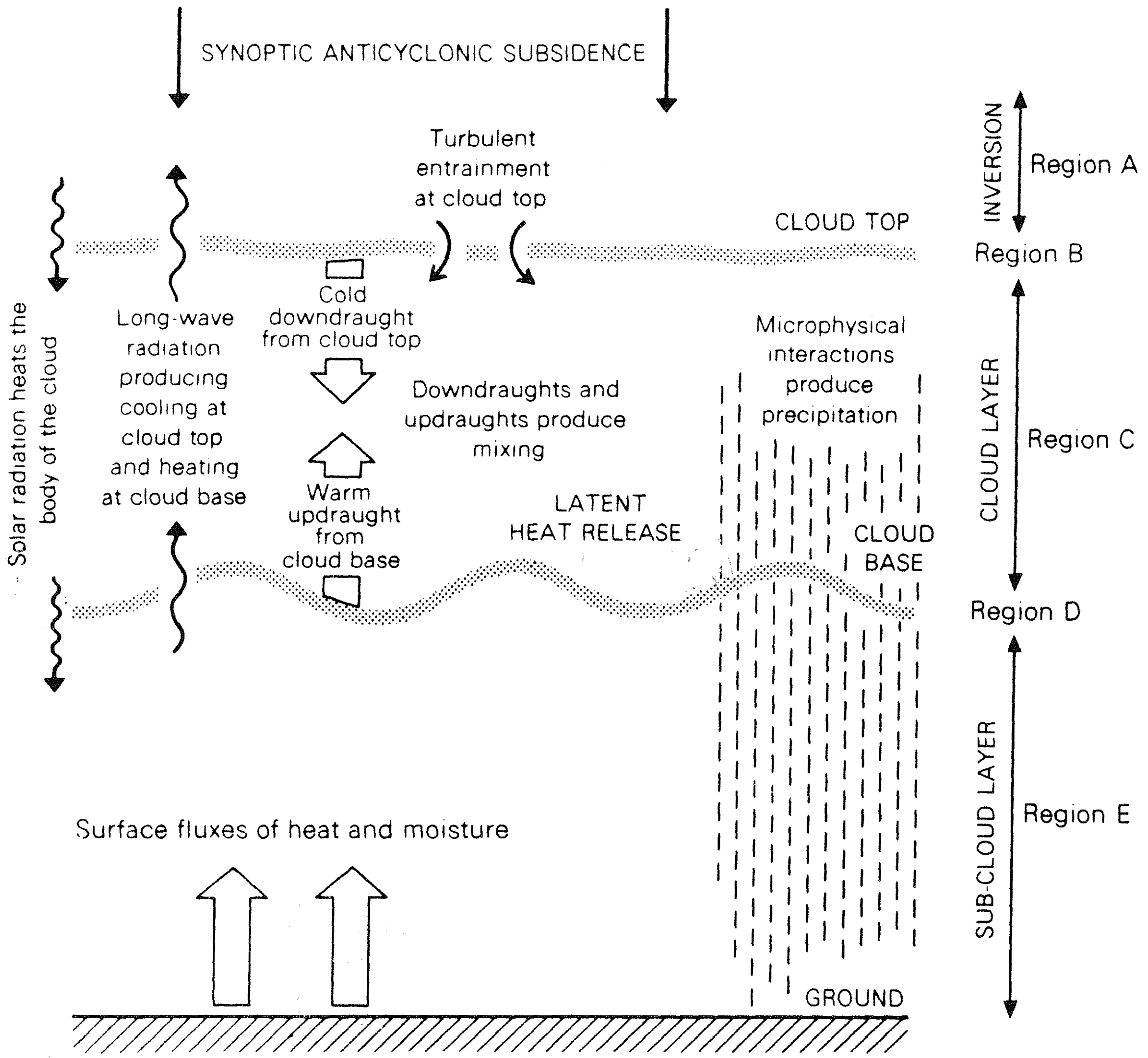

Figure 4. Summary of physical processes important to the development of stratocumulus.

shown in Fig. 5. Compared to daytime, nocturnal stratocumulus has enhanced cloud-top cooling and reduced heating (not zero as there is long-wave radiation from the ground) throughout the body of the cloud. What will happen to the cloud layer? Two competing processes may be considered (Fig. 5).

Process 1:

(i) At night there is a net cooling in the cloud layer.

(ii) Since the water vapour content remains essentially constant, cloud formation is enhanced (this has similarities to fog formation).

(iii) The net effect is that the cloud becomes denser and cloud base lowers.

Process 2:

(i) More cooling at the top and reduced heating at the base change the stability of the cloud layer leading to enhanced turbulence, principally through the presence of stronger downdraughts.

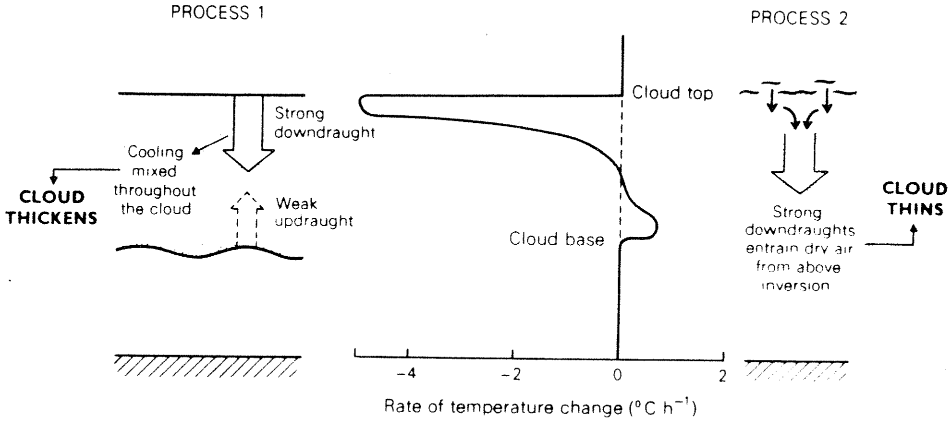

Figure 5. Outline sketch showing effect of the two competing processes on a sheet of nocturnal stratocumulus: Process 1 — radiative cooling at cloud top produces downdraughts which mix cold air throughout cloud layer and cause stratocumulus to thicken, and Process 2 — strong downdraughts entrain dry air which when mixed with cloudy air may disperse the stratocumulus.

(ii) Stronger downdraughts increase the entrainment of dry air from above the inversion.

(iii) If the air above the inversion is sufficiently dry the stratocumulus disperses through mixing.

Therefore at night there are two competing effects, one tending to thicken the stratocumulus and the other tending to disperse it. The outcome depends on the detailed structure of the cloud.

Little effort has been devoted to the investigation of Process 1 since, from a forecasting point of view, the precise details of a cloud layer are unimportant once persistence is assured. Process 2 is much more relevant since if the stratocumulus clears, surface conditions will markedly change. In fact Process 2 is the basis of James's rule (see Appendix 1). This technique was based on a statistical examination of station reports. James (1959) found two parameters to be important: D_m which measures the dryness of the air above the cloud and D_c which depends on the liquid water content and cloud thickness. If $D_m > D_c$ then the air above the cloud is dry enough to evaporate the cloudy layer completely.

James's rule is applicable over land only, mainly because the additional complications of sea surface temperature and the consequent flux of heat and moisture are too difficult to incorporate into a simple rule. It should also be remembered that James's rule is not very reliable because of the difficulty of making measurements accurate enough to distinguish between the two competing physical processes.

(b) *Daytime effects*

Consider now a sheet of nocturnal stratocumulus. As the sun rises, both the main body of the cloud and the ground warm up. At midday the radiation balance is as shown in Fig. 6. Again two processes can be envisaged.

Process 1:

(i) Both the main body of the cloud and the ground begin to warm, eventually achieving a balance.

(ii) The structure of the turbulence changes; updraughts are enhanced due to thermals, caused by the warm ground, penetrating the cloud layer from below, and downdraughts become weaker.

(iii) If the updraughts are sufficiently strong they may penetrate the inversion and induce compensating downdraughts which force dry, above-inversion air into the cloud layer.

This possibility involves a subtle balance. If the updraughts do not reach cloud top then no clearance is possible, indeed the cloud may thicken. Even if the updraughts penetrate the inversion, the air above the cloud top may not be either dry enough or warm enough to induce clearance. It is of crucial importance that the updraughts should be strong enough and the air sufficiently warm and dry to clear the cloud — or perhaps break it into small fair-weather cumulus. This balance of strength of updraught and state of the air above the inversion is reflected in Kraus's rule (see Appendix 2). Note, however, that Kraus's rule (Kraus 1943) only uses the change in temperature across the inversion. This is partly because he was more confident in measurements of temperature and partly because, in subsiding air, the hydrolapse and temperature inversion are often closely linked. Caution is necessary if the air above the cloud is unusually moist or dry.

As with James's rule, Kraus's rule is in practice less than perfect because of the delicate physical balances involved and the difficulty of making sufficiently accurate and representative measurements.

Process 2:

From Fig. 6 it is evident that sometimes the heating due to insolation and the cooling due to long-wave radiation will be of comparable magnitude within the cloud. As discussed earlier the cloud is destabilized causing turbulent motions which produce internal mixing and entrain potentially warmer air through the cloud top. The induced updraughts and downdraughts are now of equal strength and mixing is confined to the cloud layer. To put it another way, if the updraught and downdraught are of different magnitude, mixing takes place throughout the combined depth of the cloud and sub-cloud layers; if they are of equal magnitude, mixing is confined solely to the cloud layer. Thus the motions within the cloud are decoupled from those in the sub-cloud layer.

If this occurs then sub-cloud air no longer enters the cloud and the moisture supply is cut off. We have already seen that owing to entrainment at cloud top and possibly through precipitation there is a steady loss of moisture from the cloud and so if the supply is cut off, cloud base will rise and the cloud will thin and possibly disperse.

Observations show that under these conditions a weak inversion develops beneath the cloud base. If such a feature is observed during the day it suggests that there is evidence that the main cloud layer is beginning to thin and may, depending on its initial thickness, eventually clear. These ideas are discussed in much more detail in Nicholls (1984).

It is important to stress that throughout this section the discussion has concentrated on stratocumulus with radiatively driven convection. In many cases other forms of mixing (e.g. wind shear) may also be

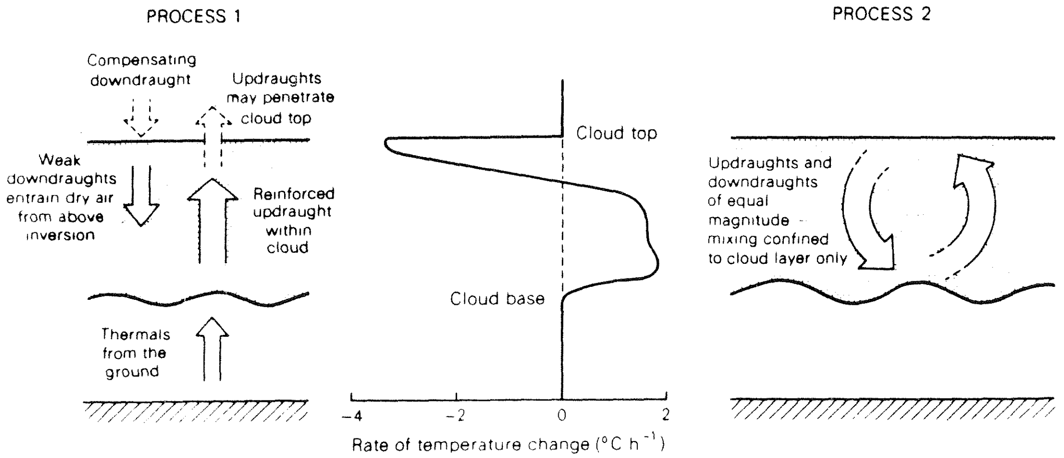

Figure 6. Outline sketch showing effect of the two competing processes on a sheet of daytime stratocumulus: Process 1 updraughts are enhanced due to warming and may penetrate the cloud top causing compensating downdraughts to entrain dry air, and Process 2 updraughts and downdraughts are of equal magnitude confining mixing to the cloud layer only.

important especially if the cloud is at low level or there are very strong winds. Then the cloud may have quite a different structure and behaviour.

5. Conclusion

This paper has attempted to summarize the main features of our current knowledge concerning the formation and dissipation of stratocumulus in simple terms. It acknowledges that stratocumulus remains a major forecasting problem, primarily because the evolution of the cloud is a response to subtle changes in the balance between a number of different, but interacting, physical processes. Forecasting rules are either completely empirical or are based on extremely simplified forms of this balance. Some progress has been made in recent years towards identifying and quantifying the important processes (see for instance the references in Nicholls, 1984). One of the aims of current research is to design numerical models which more accurately reflect these processes. Information from these models plus, it is hoped, more detailed data on an operational basis will result in better forecasts.

It is only by having a clear understanding of the processes involved that forecasters can make the fullest use of the rather meagre information available at present.

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Appendix 1 — *Nocturnal dispersal of stratocumulus over land*

James's Rule

The cloud will break if $D_m > D_c$ where D_m is the maximum depression (°C) of the dew-point below the dry-bulb temperature in the 50 mb layer above the cloud, and D_c is the value given in Table A1.

Table A1. *Values of D_c (°C)*

z (mb)	h (g kg^{-1}) 0.25	0.5	0.75	1.0	1.25	1.5
10	—	–	1.00	3.0	6.00	8.5
20	0.00	2.5	5.00	8.0	10.00	13.0
30	4.00	7.0	9.00	12.0	14.50	17.0
40	9.00	11.0	14.00	16.0	19.00	21.0
50	13.00	15.0	18.00	20.5	23.00	26.0
60	17.00	20.0	22.00	25.0	27.00	30.0
70	21.00	24.0	26.50	29.0	32.00	34.0

Where h is the difference (g kg^{-1}) between the humidity mixing ratios at the top and bottom of the 50 mb layer below the cloud, and z is the cloud thickness (mb).
Note: a linear hydrolapse in the layer is assumed.

The technique applies under the following conditions:

(i) The stratocumulus sheet is bounded at its top by a large hydrolapse, that is, a rapid decrease of humidity with height through the region of temperature increase.

(ii) There is no surface front within 400 miles of the locality of the cloud sheet.

(iii) The cloud base is above the condensation level of any convection from the sea (the rule applies only to stratocumulus over land).

(iv) The cloud sheet is extensive, covering several hundred square miles, and gives almost complete cloud cover, more than 6/8 for at least 2 consecutive hours. (The cloud was regarded as having dissipated if it broke to 2/8 or less for at least 2 consecutive hours.)

Failures of the technique in day-to-day forecasting can often be attributed to:

(i) inaccurate assessment of the cloud thickness (in the absence of reports from aircraft), and

(ii) uncertainties as to the magnitude and steepness of the temperature inversion and hydrolapse because of the lag of radiosonde elements.

Appendix 2 — *Dissipation of stratocumulus by convection*

Kraus's Rule

A cloud layer will not disperse by convective mixing with the air above if the pressure at the cloud top is less than P_c, as given below. (If the pressure at the 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 (mb) and values of P and a are given in Table AII.

Table AII. *Values of P and a (mb)*

Temperature 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	0.87	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