The haar of north-east Scotland

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SUMMARY

Case studies of three sea fog (haar) situations off the coast of NE Scotland are presented. The studies are based on aircraft and mini-sonde observations and satellite imagery, and are used to assess the roles of physical and dynamical factors in developing and advecting sea fog in order to improve guidance to forecasters at coastal outstations. Two fogs had formed 2–3 days previously west of Ireland and had been advected round the N coast of Scotland, but the second fog had lifted into stratus by the time the aircraft had reached it. The third fog was warmer than the sea and appeared to be in the development/advection stage characterized by much stronger winds and lower liquid water contents.

Analysis and some modelling of the one-dimensional heat and water budgets of sea fog suggest that after a period of development during transport over a relatively cold sea, initially ‘warm’ advection fog tends to become a diurnally modulated, self-maintaining ‘cold’ radiation fog held at a temperature 1–2 degC below sea surface temperature. In other words, radiative cooling has depressed the temperature of the fog far enough below sea surface temperature for this cooling to become balanced by latent and sensible convective heat input from the sea surface.

The aircraft observations demonstrated the convective control of the vertical structure of ‘cold’ sea fog, and the influence of diurnally forced wind systems (in this case, a heat low formed over the Scottish mainland) in moving and moulding the fog.

1. INTRODUCTION

The incursion of sea fog across the coasts of the United Kingdom occurs most frequently in spring and summer, particularly around the northern and eastern coasts of Scotland, where visibility is below 1000 m 5–10% of the time early on summer mornings. The onset of fog is often sudden and unexpected, which is a continuing cause of concern to coastal outstation forecasters and their customers—in particular the helicopter operators providing the air link with the offshore industry.

This paper describes a study of fogs (haar) in the Moray Firth area of NE Scotland using data from the Hercules aircraft of the Meteorological Research Flight (based at the Royal Aircraft Establishment, Farnborough). This was supplemented by AVHRR satellite data, conventional synoptic data and data from a portable ‘mini’ radiosonde (the Kaymont Airsonde) released from Lossiemouth (57°43’N 03°19’W) over a period bracketing the time the aircraft was flying in the area.

The main objectives of this study were: (i) To identify additional information and guidance necessary to aid the coastal outstation forecaster in his task. (ii) To gain further insight into the physical and dynamical factors controlling the development and movement of sea fog.

The first objective has been pursued by Findlater (1987) and by Taylor (1987). The latter paper demonstrates a fair simulation of one of the days studied (27 April 1984) with the Meteorological Office mesoscale model.

This paper follows the second objective. This objective is admittedly rather general, but the authors found that existing studies of sea fog in the vicinity of the United Kingdom were largely descriptive, old, and based on observations from a single location. Summaries of lighthouse observations by Buchan (1902) and Dixon (1939) provided some basic climatological data. Taylor (1917) studied sea fogs near Newfoundland and found that fog temperatures ranged from several degrees warmer than the sea to 1–2 degrees (C) colder than the sea. ‘Warm’ fogs tended to occur in warm air advection conditions, while
'cold' fogs tended to occur in winds too light to allow kite flying. Douglas (1930) suggested that when warm air advection had ceased, radiative cooling would reduce fog temperature until a saturated adiabatic lapse rate was established. He also found that the tops of such fogs exhibited an appearance similar to stratuscumulus. Pile et al. (1979) used shipboard data to infer a similar life history for sea fogs along the California coast.

Studies by Lamb (1943) and by Spink (1945) suggest that Scottish haar is often persistent and is associated with anticyclonic conditions in which inland day temperatures may exceed (foggy) coastal temperatures by 15 degC. Alexander (1964) and Harrison and Phizacklea (1985) have noted the role of tides in the estuaries of Scottish rivers in bringing fog further inland.

The rudimentary knowledge of sea fog, the lack of progress in the forecasting problem and the realization that satellite and research aircraft observations could be used to obtain some information on the three-dimensional structure of sea fog and its evolution motivated the present study. It was hoped that such a study might throw some light on some rather general (and interrelated) questions such as:
(i) What factors determine the temperature of sea fog in relation to sea temperature?
(ii) What is the heat and water balance of the fog?
(iii) What space and time scales are important?
(iv) What controls the observed vertical structure and depth of the fog?
(v) What is the nature of the interaction of physical and dynamical factors in the life cycle of the fog?
(vi) What is the influence of the diurnal cycle?

2. THE EXPERIMENT

(a) A pilot mini-sonde study

The main experiment using the MRF aircraft was preceded by a pilot experiment in 1983 in which Kaymont Airsones were released from Lossiemouth 1 km from the shore in haar conditions. It was found that:
(i) Fog appeared when screen temperature (in the early morning) fell close to sea temperature.
(ii) The fogs observed deepened at 50–100 m h⁻¹ to reach maximum depths of 150–250 m at about sunrise, thereafter decreasing in depth.
(iii) The inversion base lifted with fog top, and fog top temperature was preserved during much of the fog lifetime.

Similar rates of deepening and the preservation of fog top temperature have also been noted in land radiation fogs by Findlater (1985a), but it was not clear to what extent the observed changes in sea or land fog were due to advection or development. In addition, observations of fogs at the boundary between two thermal regimes (land and sea) are difficult to interpret.

(b) The aircraft campaign

During 1984 and 1985 haar seasons, the Hercules C-130 aircraft of the Meteorological Research Flight was flown on selected days above and within the haar. The operational, instrumental and experimental details of this aircraft have been described by Readings (1985) and Nicholls (1978).

These flights were coordinated with radiosonde observations from Lossiemouth, satellite data and supplementary coastal observations. The whole exercise was known as Project Haar, a brief description of the operational aspects of which has been given by Findlater (1985b).
Figure 1. The experimental area and aircraft flight tracks used in 1984 and 1985. Land synoptic stations are shown by dots and stationary reporting ships or platforms by open circles. RS indicates the location of the radio-sonde station at Lossiemouth.

Figure 1 shows the experimental area centred on the Moray Firth and the tracks flown by the Hercules aircraft in support of the radiosonde observations from Lossiemouth. Tracks were flown in a blunt sawtooth pattern, i.e. with level runs of three minutes each at 760 m a.m.s.l. and 150 m a.m.s.l. (minimum safety height in fog) interspersed with three-minute ascents and descents at 3.5 m s\(^{-1}\). This flight mode generated about 25 profiles through and above the fog layer on each flight. By flying around one of the triangular tracks twice profiles were obtained at the same positions after an interval of approximately 72 minutes to examine time changes.

In some areas an undulating flight was made by repeatedly moving from 60 m below fog top to 60 m above it to examine small-scale changes in the fog top height and the temperature and height of the underlying inversion base. Separate flights were made after the clearance of fog to measure sea temperatures by radiometric methods over the experimental area.

In 1984 full observations, including sea temperatures, were available from MV *Seaboard Illustrious* (GTRA), stationary at 58°18'N 03°06'W, and from the Beatrice 'A' platform, without sea temperatures, at 58°06'N 03°06'W and from land synoptic reporting stations in the area. Locations are shown in Fig. 1.

(c) **Aircraft data**

The basic meteorological parameters (temperature, dew-point and wind) were recorded continuously throughout the experimental area and are displayed at a frequency of 1 Hz in most of the composite diagrams. Some use has also been made of high frequency (20 Hz) wind data to show the turbulence characteristics. Standard correction
procedures were used to remove drift in the datum of the inertial navigation system from the data in order to ensure accurate winds. Problems with one of the wind vanes resulted in the loss of some data on 26 July 1985. Radiometric data were recorded on all flights but as a consequence of the need to make ascents and descents the resultant frequent changes in ambient conditions meant that the instruments had insufficient time to stabilize. A Barnes radiometer was used to measure fog top and sea surface temperatures. The latter were obtained either below lifted haar or on a separate flight during clear conditions within a few days of the haar flight.

Liquid water content was measured with a Johnson–Williams sensor. Fog droplet spectra were obtained with an Axially Scattering Spectrometer Probe (ASSP) on the 27 April 1984 flight and with a Forward Scattering Spectrometer Probe (FSSP) on 26 July 1985. Incorrect range settings for the ASSP on 8 June 1984 meant that very limited data were gathered within cloud and so are not included in this paper.

3. The case studies—synoptic context

The mode of presentation of a set of case studies and the associated discussion depend upon the relative qualities of the studies, and the extent of their similarities and differences. Flights were made on 27 April 1984 (H656), 8 June 1984 (H667) and 26 July 1985 (H733), and these days will be referred to as cases A, B and C respectively. Most of the observations of value were obtained on case A, and this is reflected in the space devoted to it later in this paper. The initial description of the cases is presented in parallel in order to highlight the main similarities and differences between the cases. This is followed by a more detailed attempt to interpret case A, supported by some modelling.

(a) Synoptic situation

Figure 2 shows the synoptic charts for 09 and 15 GMT for case A and for 15 GMT for cases B and C. The lifted sea fog boundary in case C is shown in Fig. 2(d) as a dashed line. The situation for cases A and B was essentially anticyclonic, but the associated wind field in each case was locally profoundly modified by the formation during the morning and early afternoon of a heat low over the Scottish mainland. Inland temperatures exceeded coastal temperatures by over 15 degC in case A and 10 degC in case B.

Cases A and B fogs had formed 2–3 days previously in the east Atlantic and were advected around the north of Scotland to ‘settle’ as established ‘cold’ fogs in the vicinity of the Moray Firth. However, there were important differences between cases A and B by the time the flights were made. The fog in case A was colder than the sea, and the surface winds were generally less than 5 m s⁻¹. However, winds in case B were about 10 m s⁻¹, the fog had lifted into stratus, the air above fog top was much drier and the air temperature appeared to be 0–5 degC above s.s.t. The latter feature may have been due to the southward advection of the fog over the belt of cold water lying across the mouth of the Moray Firth (Fig. 6(b)); case B was not thought to be a warm advection fog in the classical sense.

In contrast to cases A and B, the fog in case C formed in a strong warm SE airstream associated with a thundery low moving NE over the Irish Sea. No heat low formed over the Scottish mainland on this day because thick cloud and rain moved into the area. The case C fog was warmer than the sea and was being advected NW by surface winds approaching 15 m s⁻¹.
Figure 2. Synoptic situations for case A (a/b, 09/15 GMT on 27 April 1984); case B (c, 15 GMT on 8 June 1984); and case C (d, 15 GMT on 26 July 1985).
Figure 3. The distribution of haar on 27 April 1984. (a) At 0836 GMT by NOAA-8 satellite; (b) at 1448 GMT by NOAA-7 satellite. (Photographs by courtesy of the Electronics Laboratory, University of Dundee.)
(b) Satellite imagery

The satellite imagery presented in this paper has been processed by the Meteorological Office HERMES system (Turner et al. 1985; Eyre et al. 1984). Satellite pictures (Fig. 3) show the distribution of haar at 0836 and 1448 GMT for case A. The deep nighttime penetration of the haar up the valleys of the north-western highlands and of the southern lowlands is evident at 0836 GMT, but the haar had retreated to the sea by 1448 GMT.

In particular the well-defined clearance in the southern part of the Moray Firth in the afternoon appeared to be because of an off-shore wind developing there in association with the circulation around the heat depression over the central highlands, where afternoon temperatures exceeded 25°C and cumulus cloud based at 3 km a.m.s.l. developed. Patches of snow are also visible in the central highlands.

False colour contouring of the temperature of the fog top at 0836 and 1448 for case A showed that during the six-hour period between the two images the temperature of the fog top over the whole area has increased by 1–1·5 degC. The fog temperature at 0836 h is shown in Fig. 4.

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<td>+3·5</td>
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Figure 4. Fog top temperatures at 0836 GMT on 27 April 1984 using false colour imagery from the NOAA-8 satellite.
On the day prior to case B, sea fog lay in extensive areas to the east and north of Scotland and was moving slowly south under the influence of light northerly winds. By the afternoon of 7 June fog had cleared from much of the Moray Firth by lifting to low stratus and dispersing, but towards midnight 00 GMT (8 June) the fog had reformed in places over the sea and surrounding districts, but had lifted into stratus cloud by the time the aircraft arrived in the area.

Satellite photographs of the area taken at 1009 and 1435 GMT on 8 June show the clearance of cloud in the Moray Firth. The Hercules aircraft operated in the area between 1015 and 1400 GMT gathering data during the lifting and dispersal phase. Cloud top temperatures inferred from satellite imagery at 1007 h were fairly uniform near 2°C, decreasing slightly towards the east. By 1613 GMT, there was a steep temperature gradient in the outer Moray Firth associated with the retreating edge of the cloud sheet.

The interpretation of satellite photographs for case C was hindered by the presence of extensive high cloud, but with the assistance of ship data, it was possible to construct a line drawing (Fig. 5) showing the bodily displacement in 24 hours of the fog by about 600 km (i.e. 7 m s\(^{-1}\)) in a NNW direction.

(c) Sea surface temperature distribution

The s.s.t. distributions for the three case studies are shown in Fig. 6. Case A was based on a s.s.t. mapping flight made on 1 May 1984 (four days after case A) and a satellite image obtained at 1837 GMT on the same day. A band of warm water extends eastwards near the axis of the Moray Firth, and there is cold coastal water off the coast near Wick and Fraserburgh (Fig. 1). The s.s.t. estimates from the satellite image were 1-1.5 degC colder than those estimated from the aircraft. As the latter compared well with some local ship observations, the aircraft s.s.t. observations were taken as correct and the satellite observations used to give a better areal coverage of the s.s.t. patterns.
Figure 6. Sea surface temperature isotherms for case A; case B; case C.
In case B a belt of cold water at about 8°C extended across the mouth of the Moray Firth, and was flanked by quite large temperature gradients. There was no obvious relation between the patterns of sea surface and cloud top temperature.

It was not possible to obtain aircraft observations of sea temperature for case C, but by a fortunate coincidence the Ocean Weather Ship *Cumulus*, on passage from Rotterdam to station L via the Orkney Islands, passed through the experimental area on the morning of 26 July along track marked αβ (Fig. 6(c)) making hourly observations of weather and sea temperature. The ship sea temperatures were about 0.3 degC below the air temperature, but about 0.8 degC higher than the sea temperatures obtained in the vicinity of the line during a separate flight made on 1 August 1985. The cause of this discrepancy cannot be ascertained, so the ship temperatures were reduced by 0.8 degC before constructing the pattern of sea temperatures shown in Fig. 6(c). As in case B, there is a belt of cold water across the mouth of the Moray Firth, but less pronounced in case C than in case B.

4. THE CASE STUDIES—AIRCRAFT OBSERVATIONS

(a) Wind field and fog distribution

Figure 7 shows the distribution of fog and wind fields over the Moray Firth for case A. Figure 7(a) shows the northern limits of a clearance spreading from the Banffshire coast derived from the first satellite photograph at 0836 GMT, from the aircraft traverses.

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Figure 7. General distribution of haar on 27 April 1984. The locations of points A–F from Fig. 1 are also shown. (a) Edge of fog as determined from satellite photographs and aircraft traverses. Times are GMT. α, β indicate positions of soundings shown in Fig. 11. The rectangle indicates the area for which a more detailed kinematic analysis was made. (b) Height of fog top in metres. —— 1100–1230 GMT; ——— 1230–1400 GMT. Estimated location of fog edge at 1200 GMT is also shown. (c/d) Wind flow at 150/760 m in/above the fog layer. ← streamlines; ——— isochths, speed in m s⁻¹.
near point B at 1132, 1238 and 1350 GMT, and from the second satellite photograph at 1448 GMT. This clearance reached Lossiemouth at 1615 GMT, but was short-lived since fog returned at 1830 GMT with a general visibility of 200 m.

Figure 7(b) shows the topography of the fog top derived from aircraft and minisonde observations. This topography is reflected to some extent by the temperature distribution of the fog top (Fig. 4). A steep increase in fog depth is evident across the mouth of the Moray Firth. During the second circuit, a similar contour pattern was observed, but there was an overall decrease of fog depth of 20–70 m, particularly near point B, just ahead of the most rapidly retreating edge of the fog (Fig. 7(a)) off the Morayshire coast.

Wind fields at 150 and 760 m derived from the aircraft observations are shown in Figs. 7(c), (d). Vertical cross-sections of wind along the legs A–B–C, C–E and E–B for the first circuit of the triangle are shown in Fig. 8. The wind field had not changed significantly by the time the second circuit was flown just over an hour later (not shown).

Three main wind regimes can be distinguished:
(i) The winds in the fog layer (Fig. 7(c)) are SE over most of the sampling area, but near the Morayshire coast are weaker and backed E in association with the heat low over the central highlands.
(ii) There is a broad NE flow above about 400 m (Fig. 7(d)). This flow weakens towards the NW of the area and then reverses to become SW'ly.
(iii) In between (i) and (ii) is a complex transition zone containing marked directional wind shear, but rather light winds.

For case B, Fig. 9(a) shows the retreat eastward during the afternoon of the lifted fog which displayed a diffuse base (Fig. 9(b)). Figures 9(c), (d) show the wind field below and above the cloud layer derived from aircraft observations. Winds are generally 5–8 m s⁻¹ in the east of the area—much stronger than in case A—and, except at low levels near Lossiemouth, are generally unaffected by the heat low circulation further south. There is relatively little change of wind with height.

For case C, a time cross-section of wind and fog at Lossiemouth is shown in Fig. 10(a). With strengthening of the winds during the early morning, sea fog at Lossiemouth quickly dispersed. Soon after 07 GMT, the edge of the main fog belt (Fig. 5) reached Lossiemouth as low cloud, and lowered almost to the surface by early evening. The fog top height derived from the aircraft measurements shows (Fig. 10(b)) variations between 600 and 700 m.

The wind fields at 150 m (Fig. 10(c)) and 775 m (Fig. 10(d)) show relatively little variation in direction or speed.

(b) Vertical structure

A disadvantage of the safety limitation of the aircraft altitude to a minimum of 150 m in fog was that direct observations of air temperatures near the sea surface were not possible, but were indirectly estimated by extrapolation of the observed temperature profile to the sea surface along the saturated adiabat. The difference between s.s.t. and surface air temperature was then estimated using the s.s.t. observations presented above.

Of the many profiles measured through the fogs only a representative selection is shown here. Figures 11(a) and (b) illustrate profiles taken at the same position in case A, with a time difference of about 72 minutes, in a deep fog. Figures 11(c) and (d) show similar soundings for case A in an area where the fog was shallower. Appropriate sea temperatures are included in the diagrams.

The principal features of the vertical structure through and above the fogs were:
(i) Surface air temperatures in fog in case A were observed to be 1–1.5 degC below s.s.t.
Figure 8. Vertical cross-sections of winds along the flight tracks A–B–C, C–E, E–B. (a) Leg A–B–C; (b) Leg C–E; (c) Leg E–B. ——— isogons (degrees true); ———— isotachs (m s⁻¹); <1 m s⁻¹ dot-shaded; the fog top mixing layer between fog top and base of inversion is shown hatched.
over most of the Moray Firth but about the same as s.s.t. over coastal cold water pools near Wick and Fraserburgh (Figs. 1, 6(a)). In contrast, the fog in case B was about 0.5 degC and in case C about 1 degC warmer than the sea.

(ii) All cases exhibited a steep temperature inversion at fog top; of 10–12 degC in case A, 6–8 degC in case B and 5 degC in case C. About half the change in temperature occurred within 10–15 m. Figures 11(a) and (b) are taken from descents whilst Figs. 11(c) and (d) are from ascents in case A. There is a noticeable change of lapse rate between all such descents and ascents in the top 50–100 m of the fog. Descents show a near-isothermal layer whilst ascents show a near-saturated adiabatic lapse rate right up to the fog top. It is considered that the near-isothermal lapse rates are due to instrumental lag as the aircraft descends quickly through an interface with a drop of some 10 degC in temperature. Turbulence measurements support this view, in that there is no detectable change in turbulence between the isothermal layer and the adiabatic layer below it. This problem was also reported near the top of marine stratocumulus by Nicholls and Leighton (1986) using the same sensors.

(iii) There was a decrease in dew-point of over 10 degC above fog in case B, but little change was observed in cases A or C.

(iv) Profiles of liquid water content (l.w.c.) were observed to have slopes parallel to the adiabatic slope in cases A and B. Liquid water contents in excess of adiabatic appear in Figs. 11(c) and (d), and are interpreted as evidence that the sea fog extends to the sea surface, where its l.w.c. is probably comparable to the lateral displacement of the l.w.c.
Figure 10. General situation on 26 July 1985. (a) Time cross-section of temperature (°C), wind and cloud at Lossiemouth from 00 to 19 GMT. The period of aircraft operation on flight H733 is shown by a double line at the top of the diagram. (b) Fog top height (m) from aircraft data. --- 1029–1220 GMT; ---- 1225–1349 GMT. The 700 m contour has been emphasized to indicate changes over the period of operation. (c/d) Wind flow at 150/775 m, in fog. ↔ streamlines; ----- isotachs, speed in m s\(^{-1}\).

Figure 11. Sample profiles of aircraft data, 27 April 1984. (a) Deep fog at α at 1154 GMT. (b) Deep fog at α at 1305 GMT. (c) Shallow fog at β at 1730 GMT. (d) Shallow fog at β at 1342 GMT. The locations (α, β) of the mid points of these profiles are shown in Fig. 7(a). Each profile took about 4 minutes during which the aircraft travelled about 25 km. \(T\) = temperature; \(DALR(SALR)\) = dry(saturated) adiabatic lapse rate; \(T_d\) = dew-point; \(T_s\) = sea surface temperature; \(ALWC\) = adiabatic liquid water content. Zero offsets on some of the original liquid water content profiles have been removed.
Figure 11. (Continued)
profile to the right of the adiabatic profile based at sea surface. In case C, the l.w.c. profiles were more irregular (Fig. 12) and generally less than adiabatic in value and slope. (v) Extrapolation of the (adiabatic) l.w.c. profiles observed in case A below the lowest flight level (150 m) gave an indirect estimate of the level of fog base. This suggested that the fog base was mainly 20–60 m above the sea surface, but with the occasional dip down to -30 m, corresponding to a surface l.w.c. of 0.05–0.1 g m$^{-3}$. There appeared to be little change in these levels between the two circuits by the aircraft. (vi) The level of maximum l.w.c. appeared to coincide with the base of the inversion in all cases, and the fog top was 20–30 m above inversion base near the steepest part of the inversion.
(vii) The temperature recorded by radiometers viewing the fog from both satellite and aircraft was that at the base of the inversion, which coincided with the level of maximum liquid water content about 20 m below the fog top. A comparison of inversion base temperature measurements by different methods at three locations (Fig. 13) showed agreement to within about 1 deg C.

(viii) The increase in liquid water content with height was associated with increasing fog droplet radii with height. The analysis of the microphysical structure observed in case A is shown in Fig. 14: typically, mean radii of 7 μm occurred at 150 m above the sea but mean radii increased steadily to 12 μm near the fog top. Droplet concentrations remained sensibly constant with height at any given location, but varied horizontally across the experimental area from 150 to 200 cm⁻³. These findings accord with those of Slingo et al. (1982) and Nicholls (1984) from balloon and aircraft studies of stratocumulus cloud, but the highest liquid water content was closer to the fog top of 27 April 1984 than indicated in some of the stratocumulus studies.

Figure 14. Profiles of total fog droplet number density, droplet size spectra, and liquid water content from the FSSP data for ascent through sea fog at 1159 GMT on 27 April 1984 at 58° 07' N 01° 30' W. The droplet size spectra are shown as contours of the percentage normalized spectral density, such that at any height the sum of the values in each 1 μm interval is 100%. The liquid water content profile produced by adiabatic ascent from the surface (at saturation) is shown by a broken line (after Slingo et al. 1982).
(ix) Turbulence was experienced right up to fog top as shown by the high frequency fluctuations in the wind vector. Above fog top, the high frequency scatter is replaced by relatively smooth but laminated profiles characteristic of wind profiles in stable atmospheres. Richardson numbers within the inversion above fog top were of order unity, and it is likely that intermittent Kelvin–Helmholtz instabilities were occurring here, although flight was smooth.

5. DISCUSSION

The quality and extent of data obtained during case A has made it possible to attempt some analysis of the dynamical and physical aspects of this case.

(a) Dynamical aspects

The principal features of interest in case A from a dynamical viewpoint were the SE flow within the fog layer beneath the NE flow above (Figs. 7 and 8), and the development of the inland heat low and the effect of its associated circulation on the fog, particularly the fog clearance from the Morayshire coast between Windy Head and Lossiemouth (Figs. 2 and 5). Near midday, when the flights were made, the SE flow lay between a surface anticyclone centred near the Norwegian coast and the heat low forming over the Scottish mainland. At 850 mb, the anticyclone was centred near the N coast of Scotland, and appeared to be associated with the NE flow above the fog and its limit just E of Wick (Fig. 7).

The transition from the low-level SE flow to the upper NE flow also appears to be associated with the hydrostatic pressure gradient induced by the sloping inversion/fog top. The vector wind difference between the 150 m and 760 m levels (Fig. 15) can be compared with the field of mean temperature of this layer. The observed horizontal gradient is due partly to the sloping dome of cold air associated with fog top, and partly to an E–W temperature gradient in the subsided air above fog top. The general level of temperature gradient (about 5 K/100 km) corresponds to a geostrophic thermal wind of 8–10 m s\(^{-1}\), over the depth of the layer, which is parallel to but rather larger than the observed wind shear of 5–6 m s\(^{-1}\). This speed discrepancy may be due partly to error in

![Figure 15. Observed thermal wind and mean temperature of the 150–760 m layer. Arrows show thermal wind vector, V(760 m) − V(150 m), with length proportional to wind speed. Isotherms of mean temperature of layer are in °C. Location points as in Fig. 1.](image-url)
our estimate of the geostrophic thermal wind, but also partly to the existence of small ageostrophic wind components. Figure 7(b) implies maximum local rates of fog top descent off the Morayshire coast of about 2 cm s\(^{-1}\) in the vicinity of the observed fog clearance. Since wind speeds near fog top were generally low the pattern of local fog top descent is likely to be fairly representative of real (i.e. following the motion) fog top descent, except possibly in the vicinity of the steep fog top gradient near Peterhead, where local vertical velocities may be dominated by advection.

An independent estimate of the rate of descent of the fog top was made as follows: The small changes of wind in time and depth in the fog layer suggested that a kinematic analysis of the 150 m wind would be representative of the whole depth of the fog. A rectangular area (Fig. 7(a)) was divided into a 10 km grid and the wind components \((u, v)\) read from Fig. 7(c) used to derive fields of horizontal divergence, \(D\), from finite difference expressions.

Assuming that the decrease in fog top height, \(\Delta h\), is given by \(\Delta h = Dht\), where \(h = \) fog top height and \(t = \) time interval between circuits, the field of \(\Delta h\) so derived was compared with the observed \(\Delta h\) (Fig. 16). Although there was little correspondence between the details of the patterns, the maximum rates of descent deduced from the divergence field were of a similar order of magnitude to those observed, and both maxima occurred in the same general area of most rapid retreat of the fog edge. The rates of descent were rather larger than characteristic fog-top entrainment or anticyclonic subsidence rates and were probably associated with the heat low circulation.

The sharp gradient of fog depth off Peterhead was present during the flight, and also in the satellite images at 0836 and 1448 GMT (Fig. 3). This gradient is steeper in the afternoon than in the morning image. This appears to be a topographic effect—the Moray Firth is in the lee of Morayshire for a SE flow—possibly intensified by the development of the heat low.

The return of fog to the Morayshire coast after 18 GMT suggests that the flow regime above fog top resumed control of the fog circulation following the decay of local circulations induced by diurnal heating.

We conclude that the observed wind and temperature fields and decrease in fog top height in case A are dynamically consistent.
(b) Physical aspects

The temperature of the haarp in case A was 1–2 degC colder than the sea surface temperature. This allowed convection to develop and ensured that the physical factors controlling the vertical structure of the fog operated on a time scale of a few minutes and were therefore locally determined. This permitted the evolution of the heat and water budgets and structure of the fog to be interpreted in terms of a one-dimensional mixed layer model developed for stratocumulus by Nicholls (1984). A brief description of this model is given in the appendix. Symbols used below are defined in the appendix.

(i) Evolution of fog. The model was initiated at 12 GMT with aircraft observations made over the Moray Firth on 27 April 1984, and run with a time step of 5 min for 30 hours. Entrainment was modelled using the closure scheme proposed by Turton and Nicholls (1987). The inversion base was at 365 m, and the base of the fog (estimated from downward extrapolation of the observed liquid water content profile below 150 m) was near 40 m. The aircraft observations made near 12 h and again about 75 minutes later showed a significant drying and warming of the fog. Turton and Nicholls discuss various methods of estimating entrainment at fog top, and proposed a scheme of their own, which was adopted for the runs described below.

Three experiments were made with the model, all starting with the same initial conditions:

Experiment a. Two runs, with constant values of horizontal divergence of 0.5 and 1×10^{-5} s^{-1} respectively. Trends of fog base and top, and of $\theta_e$, $Q_T$, and $w_e$ are shown in Fig. 17(a). The solid lines are for the smaller divergence; the dashed lines for the larger divergence. The model indicated an initial rapid rise of fog base to about 150 m followed by a slow further rise to 200 m by the end of the run. $\theta_e$ showed a diurnal range of about 0.6 degC with no discernible trend, but there was a reduction of about 3% in $Q_T$ and a corresponding decrease of 30% in fog thickness. $w_e$ varied by a factor of about 2, but again with no evident longer term trend. Doubling the divergence depressed the fog layer, but had no discernible effect on other parameters.

Experiment b. A constant decrease in sea surface temperature of 2 degC/d and divergence of 1×10^{-5} s^{-1}. Following an initial rise of fog base by 100 m (Fig. 17(b)), the whole layer descended steadily at about 4 m/h with little change in layer thickness almost to the surface by the end of the run. The other parameters also decreased and the diurnal signal was almost erased.

$\theta_e$ decreased (after an initial rise) at about 1 K/d, while $Q_T$ and $w_e$ changed little. This appeared to demonstrate the control of sea temperature on fog temperature discussed in the appendix.

Experiment c. Constant sea surface temperature, but with a divergence peaking at 4×10^{-5} s^{-1} at about 14 GMT to simulate crudely the dynamical effect of the heat low which formed over the mainland and influenced the circulation over the Moray Firth. The initial decrease in fog top of 40–60 m/h generated by the model was comparable to the observed rate of decrease within the Moray Firth, although the initial model lifting of fog base was not observed. (The location of fog base was inferred by assuming an adiabatic liquid water content from fog top downwards. This led to estimates of fog base from about -50 to +60 m.) As regards the subsequent behaviour of the model, the increased diurnal amplitude of the fog base and top height (Fig. 17(c)) over that in experiment a is evident, but there was little change in the other parameters. This indicates that divergence affects the location of fog top and base, but has no appreciable thermodynamic effect, given that the other boundary conditions are unchanged. Note the steady rise of fog top due to entrainment during the (model) period of zero divergence.
Figure 17. Results of model runs with initial conditions corresponding to those observed at 12 GMT 27 April 1984. $Z$—height of fog layer (m); $\theta_e$—equivalent potential temperature ($^\circ$C); $q_T$—total water mixing ratio of mixed layer (g kg$^{-1}$); $w_e$—entrainment rate of fog top (cm s$^{-1}$). (a) Location of fog layer for horizontal divergence of $5 \times 10^{-8}$ s$^{-1}$: (-----); and $10^{-8}$ s$^{-1}$: (-----). Constant sea temperature (corresponding differences in other parameters too small to be shown). (b) Divergence $= 10^{-8}$ s$^{-1}$. Sea temperature decrease of 2 degC d$^{-1}$. (c) Diurnal pattern of horizontal divergences (in units of $10^{-9}$ s$^{-1}$) imposed on model. Constant sea temperatures.
The main features of interest resulting from these experiments are:

1. Sea surface temperature appears to exert a strong control on fog temperature.
2. A diurnal pattern of variation is superimposed on apparently very small changes in the mean heat and total water content of the fog.
3. Horizontal divergence significantly affects the location of fog top and base, but has little effect on thermodynamic parameters.

This suggests that if 'left to itself', cold sea fog will approach a diurnally modulated equilibrium which could persist almost indefinitely. The physical basis for this suggestion is discussed in terms of the heat and water budgets of the fog layer.

(ii) Heat and water budgets. In the appendix, sections a and b, it is shown from an analysis of the budget equations for heat and water used by Nicholls (1984) that feedback processes exist for the control of fog temperature and fog liquid water content which apply whether or not the fog extends to the sea surface.

The fog temperature is controlled through the temperature and water vapour jumps at sea surface. These jumps are associated with aerodynamic bulk transfer of heat and water vapour from the ocean to the atmosphere, and their magnitudes controlled by a balance between on the one hand the heat input to the mixed layer from the sea surface by eddy transfer of latent and sensible heat and net long-wave radiative transfer, from above the mixed layer by entrainment and from absorption of solar radiation within the fog, and on the other hand by long-wave radiative loss from fog top. Radiative loss from the top of an established fog depresses the temperature of the fog below sea surface temperature until balanced by heat input from the sea, the sun, and fog top entrainment.

The liquid water budget is controlled mainly by a balance between surface evaporation and precipitation. For case A, the 'equilibrium' value of liquid water content at fog top deduced from Eq. (13) in the appendix appeared to be about 0.3 g/kg (~0.4 g m^-3). This is lower than the observed values, and appears to be consistent with the slight drying out of the mixed layer by the model. The equilibrium level of liquid water content itself is increased by increased evaporation and radiative loss from fog top, and decreased by sensible heat input from surface radiative and eddy fluxes, absorption of solar radiation and fog top entrainment.

If the fog base is lifted so far that precipitation no longer reaches the sea, then this feedback control disappears and the transition from fog to cloud is, in a physical sense, complete.

(iii) Vertical structure. The relationship of the temperatures and dew-points within the fog to those above is of interest. In cases A and C, there was little change in humidity mixing ratio across fog top, whereas there was a large decrease in case B. It is possible that increased entrainment of dry air induced by increasing wind lifted and thinned the case B fog far more than in case A in a situation which had a superficial similarity to case A.

As regards the origin of the air mass, the similarity of the dew-points below and above fog may be coincidental. Sea fogs often form in deep, moist air masses advected northward in warm sectors of depressions. The lower layers of air are cooled and dried by condensation and precipitation of fog. With rising pressure, the whole air mass begins to subside and spread out. The unmodified air above the fog in case A lay about 1500 m below its condensation level, which is broadly consistent with a typical subsidence rate of 0.005 m s^-1 acting over about 3 days.

The humidity above an inversion in this type of situation is likely to be a significant factor in determining whether cloud in the mixed layer is likely to lie on the sea surface.
The vertical structure through the fog top and its associated inversion is still not well understood. Stratocumulus cloud of the type modelled by Nicholls is usually observed to have a sharp top at the base of a subsidence inversion with dry air immediately above. Some account of radiative cooling in the air immediately above cloud top is taken by Nicholls by introducing a 30 m layer in which cloud is intermittently present, and contains a radiative flux divergence of order 6 W m\(^{-2}\). In contrast, the sea fog described here has its maximum liquid water content at inversion base, and its top 20–30 m above this level. This must result in a radiative flux divergence of 20–30 W m\(^{-2}\) in the base of the inversion.

This raises the question as to whether there is an important physical difference between these two situations. It could be argued that an increase in flux divergence in the inversion means there is less radiative cooling available to drive the convection within the mixed layer, so that ‘convective’ entrainment decreases and ‘radiative’ entrainment increases, but the sum of these two components remains unchanged so that the upward growth of radiation fog into an inversion may not depend too much on the detailed structure of the fog top.

The measured radiative fog top temperature depends upon the relationship between the temperature and liquid water content profiles. The observations of radiative fog top temperature (Fig. 4) correspond closely to the direct measurement of inversion base temperature. Based on the observations that liquid water content decreases from about 0.5 g m\(^{-3}\) at inversion base to zero 20 m above inversion base, and temperature increases by about 6 K over this layer, the application of radiative transfer theory indicates that temperature as seen by a narrow-beam radiometer looking downwards in the 8–13 \(\mu m\) window should be about 1 K above the inversion base temperature. This is consistent (within the error of observations) with the results shown in Fig. 13.

Inversion base temperature will depend upon fog depth and air temperature at sea surface. Since the latter appears from Eq. (9) in the appendix to be closely controlled by sea surface temperature in an established sea fog, then the difference (\(\delta T\)) between sea surface temperature and fog top temperature should be reflected in fog depth. A comparison between Figs. 4 and 6 showed that the field of \(\delta T\) was fairly flat—suggesting an overall uniformity of fog depth—except in the vicinity of Peterhead where direct aircraft observations indicated a steeply sloping fog top. Indeed, the pattern of the fog top temperature field in Fig. 4 is reflected in the fog top contour pattern in Fig. 7(b).

It appears that satellite-derived fields of fog top temperature may give a rough indication of fog depth, but care is needed in interpretation. These fields may also be useful in indicating the possibility of icing in fog or cloud top.

6. CONCLUSIONS

The analysis of the field and model studies presented in this paper suggests the following sequence of events in a warm, moist air mass being transported over an (initially) relatively cold sea.

The air is initially cooled by contact with the cold sea. This cooling causes fog to develop as the air is cooled below its dew-point. Radiation from this fog gradually takes over the cooling of the air mass, and eventually depresses the fog temperature below the sea surface temperature. This now initiates convective and radiative heat input from the sea surface and entrainment of warm air at fog top which eventually balance the radiative loss. Radiative loss also leads to increasing liquid water content due to condensation and increasing evaporation from the sea surface until balanced by increasing precipitation (as drizzle) reaching the sea surface. Finally, this balance, or equilibrium, is diurnally modulated by the direct absorption of solar radiation within the fog.
The time scale of adjustment to equilibrium appears to be a few hours and increases with fog depth. This time scale is long compared with the convective time scale within the mixed layer, but probably too short to have much effect on the amplitude of the diurnal modulation of the fog.

Once established, it appears that this equilibrium could persist almost indefinitely in the absence of major synoptic-scale disturbances. Persistent fog and stratus cloud off the Californian coast in spring and summer may be a manifestation of this equilibrium in a climate less subject to synoptic disturbance than temperate latitudes.

In temperate latitudes, established sea fog is usually associated with anticyclonic conditions where the fog can be regarded as a passive system moved and moulded by the local winds, of which the heat low over the Scottish mainland in case A forms an excellent example.

On an operational time scale, the forecaster is concerned to forecast the movement rather than the development of fog. This study demonstrates the need to obtain observations over a wide area in order to assess the relative roles of movement and development of sea fog at given sites within the area; observations at one site yield insufficient data for useful interpretation.

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APPENDIX

Nicholls (1984) models the evolution of a cloud-topped mixed layer, using equivalent potential temperature, $\theta_e$, and total water content, $q_T$, as the principal dependent variables. These are conserved in pseudo-adiabatic motion (but allowance has to be made for precipitation) and are given by

$$\theta_e = \theta + Lq/c_p \quad q_T = q + q_L$$

where $\theta = $ potential temperature; $q = $ water vapour mass mixing ratio; $q_L = $ liquid water mass mixing ratio; $L = $ latent heat of evaporation; $c_p = $ specific heat of air at constant pressure. $\theta_e$ and $q_T$ are assumed to be maintained constant within the fog by mixing. $\theta$, $q$ and $q_L$ will vary with height in fog or cloud.

Following Nicholls, we write

$$h(d\theta_e/dt) = \theta_e + \theta_s + S + R_o - R_h \quad (1)$$

$$h(dq_T/dt) = E_o + E_h - P_o \quad (2)$$
where $h =$ mixed layer depth

$$\begin{align*}
\theta_{o,h} &= \text{eddy flux of } \theta_e \\
E_{o,h} &= \text{eddy flux of } q \\
R_{o,h} &= \text{net long-wave radiative flux}
\end{align*}$$

$S =$ absorption of solar radiation within mixed layer

$P_o =$ precipitation at surface.

The sign convention is positive for input of a parameter into the mixed layer. For entrainment, $\theta_h$ and $E_h$ will be positive if the mixed layer top jumps, $(\Delta \theta)_h$ and $(\Delta q)_h$, are positive.

Equation (1) is the total (sensible + latent) heat budget, and Eq. (2) is the water budget.

Now, some of the terms in Eqs. (1) and (2) are dependent to some extent on the jumps at sea surface of temperature, $(\Delta T)_o$, and vapour mixing ratio, $(\Delta q)_o$. We may define $T_s = T_o + (\Delta T)_o$ and $q_{sat}(T_s) = q_o + (\Delta q)_o$ where $T_s =$ sea surface temperature; $T_o$, $q_o =$ temperature and vapour mixing ratio of air close to the sea surface; and $q_{sat}(T_s) =$ saturated vapour mixing ratio at $T_s$.

(a) Total heat budget

We first look at the information provided by Eq. (1), and we assume that the fog is lying on the sea surface. We consider the diurnal variation to be smoothed out, which should be justifiable provided the terms are not too nonlinear.

We may define

$$\begin{align*}
(\theta_e)_s &= \theta_o + (\Delta \theta)_o = \theta_s + Lq_{sat}(T_s)/c_p \\
(\Delta \theta)_o &= (\Delta \theta)_o + (L/c_p)(\Delta q)_o \\
\sim (\Delta T)_o + (L/c_p)(dq_{sat}/dT)(\Delta T)_o
\end{align*}$$

(3)

The l.h. side of Eq. (1) now becomes

$$d(\theta_e)/dt = d(\theta_e)/dt - d(\Delta \theta)_o/dt$$

(4)

where $d(\theta)_e/dt =$ change in sea temperature following the mean motion of the mixed layer. Differentiating Eq. (3), we have

$$d\theta_e/dt = d(\theta_o)/dt - Xd(\Delta T)_o/dt$$

(5)

where

$$X = 1 + (L/c_p)(dq_{sat}/dT)$$

$$\sim 2 \text{ at the relevant temperature level (5–10°C).}$$

(6)

Looking now at terms on the r.h.s. of Eq. (1) in order, we may write

$$\theta_o = H_o + (L/c_p)E_o$$

$$= c_hu(\Delta T)_o + (L/c_p)c_q(\Delta q)_o$$

$$= u(\Delta T)_o[c_h + (L/c_p)c_q(dq_{sat}/dT)]$$

(7)

where $H_o =$ eddy flux of sensible heat from sea surface and $c_h$, $c_q =$ aerodynamic bulk exchange coefficients for heat and water vapour respectively.
The entrainment fluxes of heat and water vapour are also related to \((\Delta T)_o\) through their dependence on turbulence intensity which in turn is related to the eddy flux of virtual heat through Eqs. (32)–(36) of Nicholls (1984). This link is weaker during the day due to local stability modification induced by the absorption of solar radiation. Nevertheless, the model generated a variation of \(H_n\) in sympathy with \(H_o\), while the ratio \(H_n/H_o\) ranged between 1.4 and 2.0 during a 30-hour model run simulating case A (27 April 1984).

We may write

\[
\theta_h = H_n + (L/c_p)E_h = w_e[(\Delta \theta)_h + (L/c_p)(\Delta q)_h].
\]  

(8)

In the case studied, \((\Delta q)_h\) was small, so that the second term on the r.h.s. of Eq. (8) was an order of magnitude less than other terms, and is neglected in further discussion of this case. We therefore write \(H = a_0H\), where the coefficient \(a_0\) is about 1.6 for this case.

Turning to the radiative terms, it was found that although absorption of solar radiation nearly balanced long-wave radiative loss near midday, \(\mathcal{S}\) averaged over 24 hours is only about 20% of \(R_n\). \(R_o\) will approximate to the difference in black-body radiation of the sea and the fog above, given by \(4\sigma T^4(\Delta T)_o\).

We may therefore write Eq. (1) in the form

\[
d(\Delta T)_o/dt = -a_1(\Delta T)_o + a_2
\]  

(9)

where

\[
a_1 = [uc_h(1 + a_0) + uc_q(L/c_p)(dq_{sw}/dT) + 4\sigma T^3/\rho c_p]/hX
\]

\[a_2 = (R_n - \mathcal{S})/hX + dT/dt.
\]

Equation (9) shows that \((\Delta T)_o\) tends towards an equilibrium value, \((\Delta T)_m\), of \(a_1/a_2\) with a time constant of \(a_1\). Movement of the fog over colder water will simply reduce \((\Delta T)_m\).

Model values of parameters were: \(u = 3\) m/s; \(c_h = 0.0015; c_q = 0.001; T_e = 280.2\) K. Substitution of these values in Eq. (9) leads to a value for \((\Delta T)_m\) of about 2.5 K for a constant \(T_s\) and 1.5 K for \(T_s\) decreasing at 2 K d\(^{-1}\). These values are about twice those generated by the model. This discrepancy is due to the fact that the model lifted the fog into stratus about 100 m above the sea, which nearly doubled \(E_o\) and \(R_o\) so reducing \((\Delta T)_m\). For \(h ~ 200\) m, the time scale of feedback \((1/a_1)\) was about 6 hours.

This analysis suggests that the mixed layer is tending towards some equilibrium state. Equation (1) recognizes the existence of fog or cloud through the radiation terms and the assumption of saturation at the surface, but by itself contains no constraints on the depth of the mixed layer or of cloud or fog within it, or on the liquid water balance. These further constraints are provided by Eq. (2) to which we now turn.

\[(b)\] Liquid water budget

Splitting \(q_T\) and \(\theta_e\) into their component parts and specifying their values to apply to fog top at \(h\) where (in the model) \(q_L\) is a maximum, gives

1.h.s. of Eq. (1): \(d\theta_h/dt + (L/c_p)dq_h/dt\)

1.h.s. of Eq. (2): \(dq_h/dt + d(q_L)_h/dt\).

Also, since the air at \(h\) is saturated, and using the adiabatic relationship, we can write

\[
dq_h/dt = (dq_{sw}/dT)[(T_h/\theta_h)(d\theta_h/dt) + (RT_h/p_h c_p)(dp_h/dt)]
\]  

(10)
where \( R \) is the gas-constant/kg air and \( dp_h/dt \) is the rate of change of pressure at fog top. In practice, the second term on the r.h.s. of Eq. (10) is at least an order of magnitude less than the first term and is neglected. So inserting these relationships in Eqs. (1) and (2) and eliminating \( d\theta_h/dt \) yields

\[
h \dfrac{d(q_L)_h}{dt} = -P_o + \left( E_o + E_h \right)/X - \left( X - 1 \right) \left( H_o + H_h + S + R_o - R_h \right)/X(L/c_p)
\]  

(11)

where it has been assumed that \( T_h = \theta_h \).

Equation (11) shows that \( (q_L)_h \) is decreased by precipitation at the surface, convection and entrainment of sensible heat and absorption of solar radiation, and increased by evaporation, entrainment (if \( \Delta q_h > 0 \)) and long-wave radiative loss at cloud top. We now parametrize the terms on the r.h.s. of Eq. (11).

Nicholls (1984) found that within cloud, precipitation rate varied little with height and was proportional to \((q_L)_h^2\). Below cloud, evaporation decreased precipitation approximately linearly below cloud base, to zero at a distance comparable to the thickness of the cloud layer. \( P_o \) can therefore be parametrized in the form

\[
P_o = a_3(q_L)_h^2 - a_4[h - (q_L)_h/a_5]
\]  

(12)

where \( a_3 \) and \( a_4 \) are empirical coefficients, and \( q_s = \Gamma_s(q_{sat}/dT) \), \( \Gamma \), being the saturated adiabatic temperature lapse rate, and it is assumed (from the observations) that \( q_L \) follows its adiabatic value above cloud base. The second term on the r.h.s. disappears if the fog is on the sea surface.

The parametrization of the second term on the right of Eq. (11) follows that proposed in Eqs. (7)–(9), which leads to

\[
h \dfrac{d(q_L)_h}{dt} = -a_3(q_L)_h^2 + a_4[h - (q_L)_h/a_5] + a_6(\Delta T)_o X + a_7/X
\]  

(13)

where \( a_6 = [c_o u - c_h u(1 + a) - 4\sigma T^3/\rho c_p]\d(q_{sat})/dT \) and \( a_7 = (1 - X)(R_h - S) \).

Equation (13) shows that liquid water content tends towards an equilibrium value with a time scale of order \( h/a_3(q_L)_h \), which again turns out to be about 6 hours for \( h \approx 200 \text{ m} \). The equilibrium value of \( (q_L)_h \) for case A ranges from about 0.25 g/kg for \( h \approx 100 \text{ m} \) to about 0.32 g/kg for \( h \approx 300 \text{ m} \). The fog is on the surface for \( h < 160 \text{ m} \) and the second term on the right of Eq. (13) disappears.

If \( P_o = 0 \), the direct feedback mechanism ceases to operate, and an equilibrium of \( (q_L)_h \) now depends upon a balance between radiative cooling from cloud top, surface evaporation, sensible and radiative heat input from the surface and entrainment of sensible heat from the inversion above fog top. Normally, radiative loss will dominate, causing \( (q_L)_h \) to increase until the precipitation feedback begins to operate again. During the day when \( S \) is large, the cloud may begin to diminish without precipitation operating, particularly if (as is normally the case) there is a large humidity decrease above the mixed layer.

Finally, note that in comparing the coefficients \( a_1 \) and \( a_6 \), it is seen that \( E_o \) works with \( R_o \) and \( H_o \) in controlling \( (\Delta T)_o \), but against \( R_o \) and \( H_o \) in controlling \( (q_L)_h \).

Does equilibrium of \( (\Delta T)_o \) coincide with equilibrium of \( (q_L)_h \)? It appears from the above equations that if \( (\Delta T)_h \) is at its equilibrium value, there is nothing to prevent \( (q_L)_h \) taking up the equilibrium value dictated by \( (\Delta T)_o \).

It could be argued that if the external constraints of a system are made steady state, then of course the internal parameters describing the state of the system will also reach a steady state. This is certainly not true of the atmosphere as a whole. Even the statistically steady state we call climate may not be the only possible one, owing to the nonlinearity of its internal processes. However, the feedback processes identified for fog/layer-cloud
may generate one-to-one relationships between external constraints and the internal equilibrium state.

Local horizontal divergence induced by mesoscale circulations (e.g., heat lows) may be an order of magnitude higher than synoptic-scale values for short periods. The large divergence used in the model experiment (c) decreased \( h \) and \( (q_1)_h \) rapidly during the first two hours, but this effect may be considered as part of the diurnal modulation of a mean state.

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