Marine stratus clouds: Changing liquid-water and temperature structure

By J. W. TELFORD*  
Desert Research Institute, USA

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SUMMARY

Aircraft measurements with the Wyoming King Air investigated the response of marine stratus clouds off the north California coast, advected to regions with changed sea surface temperature in coastal upwelling. Cloud and associated clear-air observations were made before and near sunrise to eliminate any role of solar heating. Repeated vertical soundings followed temporal advected air. Subcloud structure was observed for the presence of moist convection, reported below clouds in some conditions. The atmospheric structure in the clear air above the marine stratus cloud layer showed shallow layers having large temperature gradients with height, apparently remaining from an earlier decoupled overlying cloud layer. Such gradients appear to be associated with thermal radiative-heat transfer in the clear air. Cloud tops were observed to be mostly warmer than the cloud just below. It is suggested that a large temperature increase over a small height difference at the cloud-top inversion may serve to neutralize radiative cooling of the cloud drops below, occurring through the atmospheric thermal radiative window.

Observations show that subcloud air is often stable to dry convection. This is not evidence that there is decoupling from the sea surface, since there is no wind shear in the layer and moisture continually moves upwards. Subcloud convection from the sea appears to continue everywhere, demonstrated by the rise of intermittent patches of condensation. Such clouds are found below the stratus deck after passage over cooler water, and are clearly responsible for the upward transport of moistures.

These transfer processes have been shown to influence cloud-drop spectra and liquid-water content in the clouds as well as the growth and dissipation of stratus clouds. It is concluded that high spatial-resolution (less than a few metres) measurements of vertical radiative-flux divergence are required to confirm the importance of the overlying warmer region in the overall heat balance of a stratiform cloud.

KEYWORDS: Aircraft Clouds Marine Mixing Stratus

1. INTRODUCTION

Marine stratus clouds have been studied by observation from aircraft on many occasions and many papers have been published describing such work. These clouds occur extensively over the eastern Pacific Ocean at temperate latitudes near the coasts of North and South America. They have also been observed near Europe and north of the British Isles where multiple cloud layers are common. The general characteristics have thus been well described, and the statistically averaged properties have been related to theoretical studies of the marine boundary layer. The paper by Driedonks and Duynkerke (1989) surveys the generally accepted approach, emphasizing averages, the supporting observations and theory, and the problems in explaining the observations. The importance of these clouds comes from their function in reflecting sunlight while producing little change in the thermal heat loss from the earth.

Many observations have been reported (such as by Wang and Albrecht (1994), carried out after sunset) but most of these ignore the sea surface temperature and its variations, so the question remains as to how much the sea surface temperature changes could be a contributing factor in cloud behaviour. An observation of particular interest concerns the apparent ‘cumulus’ clouds seen to form between the sea and the stratus

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layer, because of the information they provide about transfer processes from the sea into the air above. Several authors have described these clouds (e.g. Wang and Lenschow 1995; Boers et al. 1997; Martin and Jonas 1997) and the variations from place to place on the scale of these clouds may provide useful insight.

The clouds found below the stratus layer during these observations do not appear to be like cumulus clouds, but are flat elongated cloudy patches which may well float in place only until they accumulate enough moisture to gain the buoyancy to carry them up into the stratus cloud layer above. They are unlikely to be a precursor to a field of cumuli without the stratus, and they may be an integral part of the stratus formation and maintenance.

It is sometimes suggested that determining the conditions for a layer of stratus clouds to convert to a field of small cumulus clouds should be a goal for such studies, but there is little basis to claim that one cloud form can only develop as the outcome of the other. Frequently, the edge of stratus decks as seen in the infrared (IR) satellite photographs do not have cumulus clouds nearby. The structure of the overlying air plays an important role in cloud formation since it controls entrainment, but this overlying structure is not greatly influenced by the clouds below it and it rapidly advects across an area. Changes must originate much earlier and well upstream, and the state of the marine-boundary cloud field is not likely to have much influence. However, changes in the sea surface temperature have been shown to produce a direct response. Telford and Chai (1993b) showed that cumulus clouds lived only a short time after the stratus deck broke up upon reaching warmer water. This runaway entrainment was a result of the structure of the overlying air and the rising of warmer parcels formed at the surface. As the more buoyant parcels drove erosion upwards, the cloud water all evaporated before reaching warmer air aloft capable of limiting the buoyancy. Thus, by making parcels more buoyant at the surface, the entrainment became much larger aloft.

There seems little doubt that the substratus condensation carries moisture up into the stratus deck (Paluch and Lenschow 1991; Miller and Albrecht 1995; Wang and Lenschow 1995). Such moisture transfer also occurs in the planetary boundary layer over land, in the layers above those levels reached by dry convection next to the surface, for there the water-vapour mixing ratio increases with time (Telford 1992). Often the liquid-water content (cloud optical depth) is insufficient to give noticeable condensation in these rising parcels. Above this region, where moisture increases with height, the lapse rate is stable to cloud-free air parcels. In this paper we use the term ‘stable’ to describe a temperature, or potential temperature ($\vartheta$) profile with a lapse rate such that a parcel displaced vertically tends to return to its original level.

Such moisture transfer has been observed through the regions below stratus clouds over the ocean (Rogers and Telford 1986) and where cumulus clouds are forming over land, although it occurs sometimes without any clouds forming above (Telford 1992). This moisture transfer does not produce any temperature increase in the air at the level where the moisture increases, and occurs in the clear air where the lapse rate is between the lapse rates for dry and wet convection. The developing sequence is determined by the temperature and moisture at the surface and the temperature and moisture of the air aloft.

In seeking to explain such effects it is important that hypotheses be tested by observation, since many ideas rest on theoretical conjectures, useful because they provide a convenient basis for numerical calculation (e.g., see Moeng et al. 1995). Furthermore, in some discussions incomplete conjectures are rescued by additional conjectures—such as the idea that ‘decoupling’ from the sea surface is the result of solar heating deeper within the cloud, when the cloud top is assumed to be cooling by
thermal radiative-heat loss. A simple test is whether this effect occurs at night in the cloud layer adjacent to the surface. If the increase in solar radiation just after dawn has no observable effect, then other effects from IR radiation transfer need to be considered.

The relative role of radiative balance at stratocumulus tops in both solar and thermal wavelengths compared with evaporative effects has been a matter of some controversy. There is difficulty in conceptually defining cloud top (the geometry is complex) and the variability of optical depth in relevant wavelengths; also the direct measurement of flux divergence responsible for the net heating or cooling has proved elusive from a practical viewpoint. Thus it has been difficult to assess numerical estimates (as Moeng et al. 1995) on an appropriate scale to suggest the realism of the physical assumptions. Downward convection would thoroughly mix the air through the whole depth; continued cooling by radiation at cloud top must eventually produce cooler descending parcels throughout the layer, leading to the layer becoming cooler than the sea surface, and generate upward thermal convection, as is often postulated. However, the air is warmer than the surface. Since all such mixing also mixes horizontal momentum, shear is eliminated and with it any possibility of layers originating from advection.

The cooling of the cloud top by thermal radiative loss is a theoretical result from a calculation prone to numerical truncation errors, but a direct test of this conclusion could be made from the measurement of cloud-top temperature. If the cloud top is usually warmer than the cloud immediately below, and never cooler, then there cannot be cooling overall, and the downward convection it is presumed to engender cannot occur from this cause. The flux divergence required over 5 m or so depth near cloud top for 1 degC h⁻¹ cooling or heating rate would require a difference of about 1.5 W m⁻², between two net fluxes, where the up and down fluxes both approach 300 W m⁻². For 10% accuracy this would require individual measurements to 0.025%. There are indications that net flux (radiative-heating rate) is needed over heights of less than one metre to show what really may be happening. Measurements of the difference in total radiation could not be made one above the other from an aircraft, because one component of the radiation, either up or down, will have the aircraft in its field of view. An aircraft cannot return to the same spot relative to the cloud to the accuracy of the undulations in the ever-changing top surface, or quickly enough so that no changes can occur. The passage of the aircraft itself will completely change the top surface of the cloud. Subtracting successive measurements from the same device to cancel systematic errors is not feasible. Thus even perfectly accurate instruments offer no solution; the errors arise from the variability of the clouds. The accurate distance from the instrument to cloud-top edge could not be determined, certainly not in cloud. Hence direct measurement may not seem feasible, nor, indeed, is such a measurement needed when the temperature itself is the required result.

In the measurements described here there was a second, decoupled, cloud layer earlier in the morning, above the lower stratus cloud layer. It was clear that the cloud layer was certainly decoupled because of the large wind shear found below it at the top of the lower layer. It was seen before dawn in the IR satellite photographs but had moved away from the observational area by the time the aircraft observations began. There is no substantial shear from the surface boundary layer up to the top of the first cloud layer. This suggests that there is continuing vertical transfer of horizontal momentum, as well as the moisture which is observed to lower the cloud base, even when the layer is stable to dry convection, because θ increases with height from the surface to cloud base. In the following measurements the area chosen is over the Pacific Ocean next to the California coast, where ocean upwelling produces patches of cool water with approximately uniform temperatures but rapid transitions at their edges. This results in
changes in the cloud as it moves from the north-west, which show a direct response to the temperature of the water just traversed upwind. The abrupt large temperature changes in the water surface temperature produce large changes in the boundary layer, so that earlier changes are no longer evident, and this leaves cloud changes that are due to the water surface temperature recently crossed.

It should be noted that small changes continually occur at the boundaries of a layer, in the sea surface temperature and in the temperature of the air advecting across the top. This produces effects that seldom proceed individually to completely modify the whole layer. Hence these details, if they could be adequately observed, would show more about the time responses within the layer than its response to boundary conditions. For example, the marked difference between the up and down sounding in the same locality is due to the intermittent nature of the entrainment process, and only the integrated effects produce significant changes in the layer. The situation chosen here is relatively uncomplicated since the temperature effects are large, so hopefully providing for an unambiguous connection to be drawn between causes and effects.

The data presented here were obtained during aircraft flights off the west coast of California, near San Francisco. The flights were designed to measure the changes in the stratus clouds flowing from the north-west after a long passage over open ocean, as they encountered the changing sea surface temperatures found in the cooler upwelling areas of water near the coast. This paper emphasizes the thermodynamic variables involved in convection and entrainment mixing. The flights were begun before dawn so as to minimize the influence of heating from the sun on any thermal radiative cooling of the cloud tops.

The paper first discusses instrumental performance from which the data are derived and the thermodynamic functions used in the analysis. Referring to the data which later follow, we discuss the cloud conditions, the sea surface temperature and the wind shear. The vertical structure for each sounding is discussed in terms of how it could evolve from reversible mixing and shows that heat must be added from some additional source such as radiative transfer (latent-heat transfer of energy requires substantial precipitation which is not present).

An important new observation is that in these clouds the cloud tops are warmer than the cloud below. The transport from the sea surface below cloud brings up water and horizontal momentum and appears to result from previous latent-heat driven convection, as reported earlier. The utility of the absolute potential temperature and the moist potential temperature when discussing mixing is demonstrated, leading to some interesting questions regarding radiative-heat exchange.

2. INSTRUMENT ACCURACY

Aircraft instruments are calibrated to the best accuracy available in the laboratory but since they are used at an air speed of about 100 m s\(^{-1}\) the ultimate assessment of their accuracy depends on studying their performance in flight. Table 1 lists these calibrations for the measurements used in this paper as given by the Atmospheric Science Department at the University of Wyoming, which operates the aircraft facility. The performance of the instruments in flight determines the quality of the data collected, and examining this requires careful assessment of the measurements in terms of consistency between devices, from one altitude to another, in and out of cloud. Errors in the wind measurement resulting from heading changes require detailed discussion. Air velocity measurements require readings from several instruments, the accuracy of which can be
TABLE 1.  WYOMING KING AIR INSTRUMENTATION

<table>
<thead>
<tr>
<th>Variable</th>
<th>Instrument</th>
<th>Range</th>
<th>Accuracy</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ground velocity</td>
<td>Honeywell laser SM Inertial</td>
<td>0–4095 kt</td>
<td>13.5 ft s⁻¹</td>
<td>0.0039 kt</td>
</tr>
<tr>
<td></td>
<td>Reference System</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pitch angle</td>
<td>Honeywell laser SM Inertial</td>
<td>0–90°</td>
<td>0.05°</td>
<td>0.000172°</td>
</tr>
<tr>
<td></td>
<td>Reference System</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Roll angle</td>
<td>Honeywell laser SM Inertial</td>
<td>0–180°</td>
<td>0.05°</td>
<td>0.000172°</td>
</tr>
<tr>
<td></td>
<td>Reference System</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Heading angle</td>
<td>Honeywell laser SM Inertial</td>
<td>0–180°</td>
<td>0.02°</td>
<td>0.000172°</td>
</tr>
<tr>
<td></td>
<td>Reference System</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Attack angle</td>
<td>Rosemont 858AJ/831CPX</td>
<td>±150.0°</td>
<td>0.20°</td>
<td>0.003750°</td>
</tr>
<tr>
<td>Slip angle</td>
<td>Rosemont 858AJ/831CPX</td>
<td>±150.0°</td>
<td>0.20°</td>
<td>0.003750°</td>
</tr>
<tr>
<td>Pitot/static pressure</td>
<td>Rosemont 831CPX</td>
<td>0–85 mb</td>
<td>0.2 mb</td>
<td>0.005 mb</td>
</tr>
<tr>
<td>Static pressure</td>
<td>Rosemont 1201FA1B1A</td>
<td>0–1034 mb</td>
<td>0.5 mb</td>
<td>0.06 mb</td>
</tr>
<tr>
<td>Global positioning system</td>
<td>Trimble 2000 GPS</td>
<td>0–90° latitude</td>
<td>0.8 mm h⁻¹</td>
<td>0.000172°</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0–180° longitude</td>
<td>1.66 mm h⁻¹</td>
<td></td>
</tr>
<tr>
<td>Geometric altitude</td>
<td>King KRA 405</td>
<td>0–2000 ft</td>
<td>3% &lt; 500 ft</td>
<td>0.48 ft</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>5% &gt; 500 ft</td>
<td></td>
</tr>
<tr>
<td>Air temperature</td>
<td>Reverse flow (Minco Element)</td>
<td>–50 to +50 °C</td>
<td>0.5 °C</td>
<td>0.005 °C</td>
</tr>
<tr>
<td>Dew-point temperature</td>
<td>Cambridge 137C3</td>
<td>–50 to +50 °C</td>
<td>1.0 °C above 0 °C</td>
<td>0.006 °C</td>
</tr>
<tr>
<td>Radiative thermometer</td>
<td>Heinmann KT1985 (9.5–11.5 μm)</td>
<td>–50 to +400 °C</td>
<td>0.5 °C</td>
<td>0.1 °C</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(10 s response)</td>
<td></td>
</tr>
<tr>
<td>Cloud-droplet spectra</td>
<td>Particle measuring systems FSSP</td>
<td>0.5–7.5 μm</td>
<td>not specified</td>
<td>0.5–3.0 μm</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.0–15 μm</td>
<td></td>
<td>depending on</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.0–30 μm</td>
<td></td>
<td>selected range</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3.0–45 μm</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

FSSP is the Forward Scattering Spectrometer Probe.

referred to the inertial platform as the aircraft changes speed, banks, climbs in spiral flight, and executes pitching motion.

The air-motion measuring system is based on an inertial platform in the aircraft to provide a velocity vector for the aircraft motion relative to the earth, in an aircraft coordinate system. Thus the measurement of the vector motion of the approaching airstream from the aircraft is in the same coordinates; the aircraft motion can be subtracted to remove it and so give the air-motion vector relative to the earth. The pitot/static pressure together with the attack and slip angles must be correctly calibrated (which can be done in flight trials), otherwise the wind vector will vary as the aircraft turns. Thus, if the recorded wind does not vary with heading during circular or spiral flight, we can be assured that the whole system is accurate to the accuracy of the inertial platform. This in turn has an accuracy guaranteed by the position information it must provide (by integrating the velocity) to accomplish its navigational function. A description of these requirements can be found in Telford et al. (1977) for a more elaborate system designed to avoid the effects of aircraft flexure in strong turbulence, when precise, rapid-response, vertical velocity is needed for heat-flux estimation. In
this present case the turbulence is very low and vertical velocity is not used, but the discussion of turning flight is applicable.

The inertial platform should tilt to keep exactly level as it moves around the surface of the earth. This ensures that the horizontal acceleration needed to produce the position information does not respond to gravity. Errors in the tilt introduce gravity into the horizontal acceleration, so the platform tilts as the false position develops and the platform rocks back and forth in a pendulum-like motion. All errors propagate with an oscillating period of a pendulum of length equal to the earth’s radius; this period

\[ T = 2\pi \left( \text{length/\text{gravity}} \right)^{1/2} = 84 \text{ minutes}. \]

Since, however, a position error of six nautical miles, which exceeds any expected platform error, amounts to 0.1° error in tilt, this is negligible in air-motion estimation at 100 m s\(^{-1}\). Hence there are a large number of error sources. While heading errors are more complex, the result is similar. If the inertial velocity is continuously adjusted post flight so that its integral matches the Global Positioning System position, the resulting errors in air motion are sufficiently small for our purpose.

The errors in air speed (using air density), slip and attack angles consist, approximately, of zero offset and scale-factor errors around the operating point, and can be calibrated in flight as mentioned above. To avoid the changes in angle of the stream tubes of the approaching airstream as the lift of the wing changes, the wing circulation in turns (or gusts), the pitot/static head and the slip and attack devices are mounted ahead of the aircraft nose on a boom. Wind velocity errors from all these sources result in changes with heading during turning flight, so if there are no changes correlated with heading it is a good indication that the wind measurement is trustworthy. The wind magnitude for soundings B and C (locations are shown in Fig. 3) have been plotted in Fig. 1 to show variations with aircraft heading and altitude during spiral, constant-bank flight. Sounding B follows passage over cooler water, so with the mechanisms discussed later in the paper fairly uniform mixing should be expected. There are no obvious changes with heading in Fig. 1, although at the lowest and highest altitudes (towards the surface and cloud top) the wind is reduced.

Sounding C is located over cooler water just beyond a sustained passage over warm water. Thus convection should be temporarily halted until moist patches accumulate again at the surface, and with it vertical transport of horizontal momentum. Thus the reduced wind speed at the lowest level is to be expected, as shown in Fig. 1. Otherwise there do not seem to be any consistent changes with heading.

Thus the airflow below cloud tops down to 150 m has no significant differential shear capable of advecting changes in the vertical structure. In strong thermal convection the horizontal air velocity is almost constant with height and the variability seen here, it is suggested, is likely due to the horizontal momentum being convected upwards by irregular patches of cloudy air from the water surface. This allows restricted horizontal turbulence to develop between rising elements in the otherwise stable air. The lower altitude was set by the safety requirement for flying in complete darkness at night. The pressure altitude was plotted against radar (geometric) altitude and showed a good match, with a surface pressure given as 1014 hPa. This surface pressure was used in computing absolute and moist potential temperature.

The radiation surface temperatures agree roughly with the extrapolated surface air temperature. No correction for absorption was applied but it should be about the same from place to place, and the changes from one sea surface area to the next are the only required data in this paper. For these reasons accurate sea surface temperature is not needed.
The air and dew-point temperature are used to examine mixing at the transition from cloud top to the drop-free air immediately above, and in every calculation of the absolute and moist potential temperatures. The prescribed accuracy offers little useful information here, because it is the tracking of dew-point temperature with the air temperature that is more important. Furthermore, the response times of the instruments are not addressed in the accuracy table and need to be considered in flight conditions to verify their adequacy.

There is concern that the measurement of dew-point temperature may not be reliable within cloud, and at the cloud-top transition. Cloud water may accumulate in an instrument and affect the dew-point on leaving the cloud top and on entering the cloud from above. On leaving the cloud from below, the subsaturated air may require the mirror of a dew-point instrument to cool suddenly so that the thermal feedback control cannot follow, and there may be some water on the walls of the cavity which must first evaporate. A higher than realistic dew-point temperature is recorded for a brief period in the dry air. On entering the cloud from above, the mirror is too cold and rapid condensation wets the mirror to the extent that the heating overshoots in temperature until the servo regains control.

The scale factors in the calibration of the air-temperature thermometer and of the thermometer in the dew-point instrument are likely to be sufficiently accurate, since these can cover a wide temperature range during calibration in the laboratory and are not much affected by the air-speed influences occurring in flight. Comparison of these two temperatures in cloud verify that the offset errors are acceptable. The dew-point temperature gives the water vapour pressure from the standard table, and quite independently, the temperature and pressure give the in-cloud vapour pressure (within a fraction of a per cent of saturation with these very low updraughts). The ratio of these two vapour pressures gives the relative humidity, close to 100% in cloud. Figures 2(a)–(c), where soundings B-down, C-down and D-down are plotted, show how well this is achieved in these data. The tracking of the mixing ratio with the temperature just above clouds (one sample per second) confirms that the time constants are adequate.
Figure 2. Sounding (a) B-down, (b) C-down and (c) D-down. These show the observed relative humidity to be very close to 100% in cloud as it should be if the temperature and the dew-point temperature observations are both accurate. The cloud is indicated by the presence of liquid water; (a) shows how the mixing ratio and the temperature both increase in cloud as the top is approached. (b) and (c) show examples with expanded altitude scales where both cases have warmer layers just above cloud top. In (b) there are still cloud drops in these warmer layers whereas in (c) the drops have evaporated. In both these cases the mixing ratio is higher than in the cloud below which cannot occur by mixing and appears to require radiative heating (see text). The scale along the bottom axis is: temperature, 0 to +10 degC; vapour mixing ratio, 0 to 10 g kg\(^{-1}\); liquid-water mixing ratio, 0 to 1.0 g kg\(^{-1}\) (×10); and relative humidity, 0 to 100% (×10\(^{-1}\)).
The air-temperature instrument has a time constant of about 0.1 s whereas the dew-point measurement is much slower (seconds). Thus the almost perfect match is a little surprising since the dew-point device is considered slow to respond, and it is mounted on the aircraft skin back towards the tail where the fuselage is converging. It implies that good ventilation is achieved there, perhaps due to flow separation and the resulting local boundary-layer turbulence. Both instruments must be providing self-consistent temperatures, confirming that the accuracy is sufficient for the type of analysis undertaken here.

There is no reason to question that the observed temperatures at cloud tops are higher than the temperatures below, or that the absolute and moist potential temperatures derived from the dew-point temperature, air temperature and pressure should be in any doubt to a fraction of a degree.

3. ANALYSIS TECHNIQUES

(a) Cloud-top transition

Sounding B-down (Fig. 2(a)) confirms that the two independent measurements, the temperature and the dew-point temperature, giving the in-cloud relative humidity near 100% (±1%), are both functioning correctly. There is a slight supersaturation of a few per cent relative humidity right at the cloud-top transition but, since the dew-point temperature was only recorded once per second, the cause of this cannot be resolved. It may be due to the dew-point device, but intermittent cloud or some other effect at the cloud-surface transition may be a contributing factor.

Immediately above the cloud top, in the sounding C-down, there are very small isolated patches of cloud which are warmer than the cloud underneath as shown in
Fig. 2(b). It takes about 5 s in the sounding from the first drops to the main cloud, and the aircraft moves forward over 400 m horizontally in this time, so these patches of cloud are not one above the other. The cloud top is warmer than deeper in the cloud, and the isolated patches of drops above are warmer still.

In the case of sounding D-down (Fig. 2(c)) there are two layers of clear air just above the cloud top with a vapour mixing ratio higher than in the cloud itself, as also occurred in the isolated cloud patches of Fig. 2(b). Mixing of cloudy air with air of a lower or equal moist potential temperature from higher up, produces just-saturated air when the mass-weighted averaged mixture has the same temperature as the cloud. Thus, clear air with a higher vapour pressure or mixing ratio cannot be produced by further addition of more subsaturated air from above with a lower mixing ratio. A cloudy warmer layer at cloud top can form with radiative heating as discussed earlier and further addition of cloudy air. A higher vapour pressure in clear air can then result as further heat is added to the air by radiative transfer so that the cloud drops evaporate. As can be seen later in Fig. 6, the moist potential temperature just above cloud becomes much higher than elsewhere as a result.

Figures 2(a)–(c) all show the substantial increase in both temperature and dew-point temperature in cloud near the cloud top, compared to the cloud just underneath. Since the relative humidity remains close to 100%, the reality of the temperature increase appears to be beyond dispute. The relative humidity decreases abruptly just above cloud as expected. This temperature increase in the top cloud layers is discussed in more detail later, both in terms of the observations and their possible explanation.

(b) Potential temperatures

The absolute potential temperature is the temperature of a parcel compressed, adiabatically, to a standard pressure, usually 1000 hPa. In this case we have chosen a surface pressure of 1014 hPa, obtained from the radar altimeter, and the actual pressure (see Telford and Chai (1993a) for definitions of absolute and moist potential temperatures). Thus, when the surface pressure is used as a reference, this absolute potential temperature can be compared directly to the sea surface temperature. In a well-mixed layer it is constant with height since it is constant for the same parcel at different pressures, and hence gives information about mixing. Since it includes the latent and specific heat of the cloud drops and the specific heat of the water vapour, it remains constant in a cloudy, well-mixed layer throughout cloud and below.

The moist potential temperature is helpful where the cloud top mixes with the overlying subsaturated air: it is the wet-bulb temperature of the sample with the absolute potential temperature at the chosen reference pressure. Since the air is saturated by evaporating liquid water added with the temperature of the final result, it is independent of the initial pressure of the sample and is constant for the same parcel at different pressures, if no heat is added or removed (as, for example, by radiation). The wet-bulb potential temperature is usually defined by compressing the air and keeping it saturated by evaporation of liquid water supplied at the same increasing temperature of the compressing air. Hence the total specific heat introduced into the system by this water used to saturate the air depends on the starting temperature and pressure. Otherwise the moist potential temperature and the wet-bulb potential temperatures are very similar. Two parcels with the same wet-bulb temperatures at a given pressure have almost the same moist potential temperatures. They are used here to be consistent with the previous definitions.
(c) **Mixing at the transition layer at cloud top**

When cloudy air and subsaturated air are mixed, the temperature of the mixture results from the evaporation of the cloud drops and the different starting temperatures of the samples. At cloud top the overlying subsaturated air is 10 degC or more warmer, so a little cloudy air in the mixture cools the dry air both because of the lower temperature of the cloud and the cooling from the evaporating drops (Rogers and Telford 1986). Repeated addition of small quantities of cloud will repeatedly further cool the subsaturated mixture until saturation is reached. All stratus clouds erode upwards until they reach overlying air with a moist potential temperature high enough that mixtures are not cool enough to descend in cloud. When the overlying air maintains a moist potential temperature lower than this, subsaturated air enters the cloud so rapidly that cumulus clouds form with cloud-free air between them, providing the moisture supplied from the surface is enough to maintain cloud.

After saturation, adding more cloudy air does not produce any more evaporation, so the temperature is then averaged, and if the mixture is still warmer than the cloud it will remain warmer. If the mixture at saturation is cooler than the cloud, adding further cloud to the saturated mixture will slightly warm it. Thus, when the overlying moist potential temperature is less than in the cloud, mixtures which have enough cloud water to just saturate the subsaturated air give the lowest temperature which can be reached by varying the proportions in the mixture. Such an optimum mixture is cooler and denser than the original cloud.

If the water drops are separated and allowed to evaporate into the subsaturated air, it cools to close to its wet-bulb temperature. Since two parcels with the same wet-bulb temperature at a given pressure have the same moist potential temperature, a parcel of subsaturated air with the same moist potential temperature as the cloud top can cool to the same temperature as the cloud top when the mixture is just saturated.

When the moist potential temperature of the unsaturated air contributing to the mixture is greater than that of the cloud, mixtures will always be warmer than the cloud, and such mixtures will float on top of unmixed air. Some such mixtures will contain cloud and will be slightly warmer than the original unmixed cloud and should also float at cloud top. In this case, entrainment will be very small since further mixing cannot produce cooler mixed parcels able to descend.

Buoyant cloudy layers may be able to form on top even when cloudy mixtures are cooler than the original cloud, which mostly results in mixed parcels just sinking through the cloud. Clear-air mixtures with excess subsaturated air can be warmer than the cloud and so remain on top. Additional heating of the clear air just above cloud top, by any means such as radiation, can create a warmer, very thin, layer which cannot be cooled to less than the cloud-top temperature by further mixing. The addition of more cloudy air will give warmer cloud layers, which will not sink. Entrainment events that penetrate through such a layer can continue to give negatively buoyant parcels. The stirring produced by the cooled mixtures as they start to descend will promote more mixing. There always seem to be a few drier parcels observed descending through the cloud layer.

Entrainment is driven by both local shear and buoyancy. Intermittent turbulence by shear may occasionally displace the buoyant overlying layers to bring drier air with a lower moist potential temperature into cooler cloud just below the top, where the cooler mixtures are negatively buoyant and hence can descend.

Thus, a buoyant cloudy cap may form even when entrainment is active if the clear-air mixtures floating just above the cloud are heated additionally before further mixing.
The data show that most cloud tops sampled here are warmer than the cloud beneath (see section 6). This is difficult to explain except in this way. Radiative heating and cooling, which may occur in the clear subsaturated air where the vertical temperature gradients are very large, as found just above cloud top, would provide an explanation.

Let us consider how much cloudy air must be mixed into a sample of clear air from above to cool it to cloud temperature. Stratus clouds form when the mixing rate of air from above cloud is small compared to the mixing where cloud evaporates as fast as it forms, which gives cumulus clouds. Thus the inversion rises until air with about the same moist potential temperature is met at the inversion. Assuming air with the same moist potential temperature, consider clear air from near 1100 m altitude in Fig. 2(a), where the vapour mixing ratio is 3 g kg\(^{-1}\). To match the cloud vapour mixing ratio, cloud water must be mixed in until the vapour mixing ratio reaches 8 g kg\(^{-1}\). Since the maximum liquid-water mixing ratio is about 0.4 g kg\(^{-1}\) in cloud, for a 1 kg sample of the subsaturated air 5 g of liquid water are required. This requires about 12 kg of cloudy air. Thus mixtures with more than 8% of this subsaturated air will give cloud-free air warmer than the cloud, but with a much higher mixing ratio.

To reiterate, there can be some variability in the moist potential temperature of the subsaturated air entrained at cloud top. The clear stable layers floating just above cloud will be almost as cool as the cloud itself, and so exhibit the temperature difference with the air above, which may allow thermal radiation transfer down from the clear, much warmer, air from the inversion. Such heating would warm the air, forming new mixed cloud-top layers. Continued mixing into the cloud from inversion air with a lower moist potential temperature will give cooler cloudy mixtures which descend through the cloud. It may also form cloudy parcels warmer than the cloud underneath, as these observations show, if the clear air very close to the cloud top is warmed before mixing is completed.

This mixing process will erode the overlying air, so the cloud extends upwards. Whenever there is enough moisture to maintain a cloud, the cloud top will rise until it reaches overlying air with the same or higher moist potential temperature. If the overlying atmosphere does not provide such air, the continued rapid entrainment will lead to the formation of cumulus clouds and then to clear skies (Telford and Chai 1993b). The total evaporation of a cloudy layer leads to the cloud being replaced by a stable layer with an increasing absolute potential temperature with height, but with a constant moist potential temperature (Telford and Chai 1984) as often seen over the Pacific Ocean (Telford and Keck 1988).

4. THE OVERALL CLOUD CONDITIONS

The data discussed here were gathered on 11 September 1996. The flight was made parallel to the coast with the outgoing path at 170 m altitude, below cloud, to measure the sea surface temperature (radiative temperature at about 10 μm). The aircraft took off from Livermore, CA airport at 4 h local time (1100 GMT) and by about 5 h 30 min reached the northern end of the track. Here it ascended from 150 m to 3000 m altitude, returned to 150 m and then climbed back to 1000 m, just above cloud top; no higher clouds were encountered. The next sounding position was reached by cruising above the clouds. Sunrise was observed during the second sounding, at about 6 h 30 min. Higher clouds were visible to the north but no other overlying clouds could be seen above the flight track where soundings were obtained. The returning track kept above the outgoing flight track (Fig. 3).

Figures 4(a) and (b) show two GOES 9 IR satellite photographs with contrast enhanced to differentiate between open sea, low stratus cloud and the higher cloud layer.
They were taken at 6 h, the time the aircraft was at the north end of the flight path, and 4 h local time, respectively. The higher cloud was drifting away from the flight area towards the north-east, and was probably entraining subsaturated air and evaporating slowly at its trailing edge. Earlier, before dawn, the cloud was further to the south as can be seen in the second photograph.

The flight trajectory of the Wyoming University King Air aircraft is also shown in Fig. 4(a), superimposed on the GOES 9 satellite position above the earth, with the wind vectors added. The cloud layer is between the lower outgoing track and the return path above it and the outline of the higher cloud nearest to the flight position has been added. This diagram is in the same coordinate system as the satellite photograph. The soundings to 3000 m can be seen and the following figures analyse these soundings from the north end back to the last one at the southern end, before the return to the airport at Livermore. The winds above cloud top are given to the left of the flight track and the wind directions and magnitude in and below cloud are shown between the flight track and the coastline.

5. Cloud conditions and the subcloud transfer

(a) Variations with altitude in the soundings, after passage over sea surfaces at different temperatures

Figure 5 shows the section of each sounding measured during the climb. The sea surface radiometric temperature (from 10 μm IR emission) is given to relate the soundings to the temperature of the surface over which the air has just passed. At the top of each sounding is the local time when the aircraft climbed through cloud top at 1000 m; the upwind end lies to the north (Fig. 4(a)). The sounding positions are marked where
Figure 4. Enhanced contrast satellite photographs of the flight area. The marine stratus cloud is shown in grey, the open ocean in black and the higher clouds in white. The coastline is drawn in black. (a) 6 h local time, 11 September 1996. The aircraft flight track is superimposed; at this time it was near its northern edge. The flight path of the aircraft is displayed in normal mapping coordinates, with the five soundings A, B, C, D and E. Vectors representing instantaneous wind measurements are plotted, below cloud to the right of the track and above cloud to the left. The winds are roughly constant below the cloud tops, and in the clear air above. A wind vector magnitude is labelled to give a reference scale. (b) 4 h local time, with later flight track at 6 h superimposed. The higher cloud is moving to the north-east and has departed from the area when the measurements were obtained.
the altitude changes. The vertical transitions of the radiometric temperatures change from cloud-top temperature to the surface temperature. The soundings through the cloud show changes corresponding to the previous passage of the cloud over water surfaces to the north-west of the actual sounding position, discussed in greater detail later. The radiometric temperature measured to the north along the subcloud flight track gives the surface water temperatures actually encountered. Soundings B and D show effects of a cooler water surface, D being the more advanced form from a lower sea surface temperature and a larger area of cooler sea surface. Liquid water in substratus cloud patches appears in sounding D, and the forward-looking video camera shows flat patches of condensation below the stratus over the cooler water in sounding B. Soundings C and E show the effects of encountering warmer water. Sounding E reflects the development from the previous state of the cloud and subcloud shown in D.

(b) Controlling factors

The sequence of events during modification of the clouds as they move over the changes in water temperature may be described as follows:

Since there is little information upstream of sounding A let us start with sounding B. Just downwind of A (at 15 °C) the surface sea water becomes cooler, being at a little over 13 °C for most of the distance to B. At B the liquid-water profile in the sounding through the cloud is close to the wet adiabatic value, and without the sharp dips in liquid-water value, or the marked reduction near cloud top or at cloud base which characterize recent entrainment of subsaturated air from above. It seems possible that the reduced entrainment is because the parcels leaving the cooler water remain in the cloud, and
are not buoyant enough to push through the cloud top into the inversion air and so stir in subsaturated parcels from above. At lower levels in the cloud the continuing flux of water from the surface mixes throughout the cloud and builds the cloud base downward. The adiabatic liquid-water cloud profile in the vertical is the direct result of being well mixed.

It should be noted that small changes continually occur at the boundaries of a layer, in the sea surface temperature and in the temperature of the air advecting across the top. This produces effects which seldom proceed individually to completely modify the whole layer. Hence these details, if they could be adequately observed, would show more about the time responses within the layer than its response to boundary conditions. The situation chosen here is relatively uncomplicated since the temperature effects are large, providing for an unambiguous connection to be drawn between cause and effect. The marked difference in water mixing ratio between up and down soundings in the same locality, for example, is the result of the entrainment process, which acts intermittently at cloud top. Such individual variations do not seem to be significant; clouds are modified by the overall accumulation of such effects.

Below cloud the absolute potential temperature increases from the sea up to cloud base, so this layer does not sustain dry convection. The water substance transfer is thus driven by latent heat, as described in Telford (1992). The videotape from the flight shows stratus patches of condensation, elongated clouds resembling small patches of stratus cloud, close to the sea and at various heights below the recorded stratus cloud base. The aircraft did not pass through any of these cloud patches during this sounding.

The sea surface temperature is higher from sounding B along to C, at 15 °C, and the warmer air parcels from the surface produce changes in the cloud liquid-water profile. As sounding C develops, the parcels leaving the surface with the higher buoyancy travel up through the cloud and then encounter the subsaturated inversion air and mix some of it back into the cloud. This subsaturated air dilutes the liquid water near the cloud top. Some of the blobs are negatively buoyant and sink through the cloud and erode the cloud base upwards. This happens as this drier air, with the higher condensation level formed by mixing cloud air with subsaturated air from above, replaces the bottom layer of the cloud. (Telford and Keck 1988). The centre of the cloud has a lower absolute potential temperature than the newly modified cloud both above and below the centre, since less warm overlying air has been mixed into this level.

The passage to the next sounding at D is across a second region of cooler water, of greater extent and lower temperature than that before sounding B. The result is that substantial cloud was encountered below the general stratus cloud base. (The lower temperature of the parcels leaving the sea surface results in less buoyancy and hence these parcels accumulate below cloud.) The liquid-water content of the stratus cloud layer now shows less than complete vertical mixing, and an increase in liquid-water content, greater than the adiabatic (well-mixed) value appropriate for the cloud higher up, towards cloud base. It appears that the parcels from the surface come to rest lower and lower in the cloud, without rising up to cloud top as appeared to occur in sounding B where the water was not quite so cool. Thus the excess water builds the stratus cloud base downwards. Rogers and Telford (1986) previously reported this behaviour.

The temperature of the surface water now increases considerably. At sounding E the increased buoyancy of the parcels leaving the surface has swept the subcloud condensation up into the middle of the stratus layer, following the adiabatic profile to give a higher liquid-water mixing ratio than occurs elsewhere in these soundings. The highest liquid-water content has the lowest absolute potential temperature in the cloud, equal to that of the subcloud convection in the previous sounding, apparently the
source of this air. More buoyant parcels have also reached cloud top and begun to stir in subsaturated overlying air, and the cloud base is depleted in water content, just as happened in sounding B for the same reason.

(c) Structure of convection from buoyant condensation below cloud limited by entrainment of subsaturated air

There now remains the interesting question as to why the moist potential temperature is almost constant from the lowest levels to cloud top. The absolute potential temperature increases upwards below cloud in all cases, indicating the lack of dry convection when associated with a cooler sea surface.

The question is: how do parcels with condensed liquid water modify the layers through which they rise when the continuous mixing on the way up with the entrained subsaturated surroundings evaporates their liquid water? Such condensing parcels can rise whenever the environment is stable to thermal convection but less stable than the wet adiabat. In the case of thermal convection the rising warmer parcels always remain buoyant once the environment is established by the return flow. The environment thus becomes well mixed. When there is condensation, however, entrainment dilution of the parcel can reduce the parcel to below saturation and so stop the parcel rising, and it can come to rest at any height.

From the lowest point near the surface up to cloud top, increasing absolute potential temperature shows that this region is stable to dry convection. The almost constant, or slightly decreasing, moist potential temperature is probably caused by a similar process to that which tends to leave a cloudy layer with a constant moist potential temperature, after it has evaporated following the mixing in of subsaturated air from above (Telford and Chai 1984). Moist parcels with drops can proceed up or down on the wet adiabat as long as they contain drops, but once mixing has evaporated all the drops they cannot move up or down further because they are now dry parcels in an environment stable to dry motion.

If a cloud layer topped by an inversion is evaporated by entrainment from above, there are two sources of air—the feeder supply from the surface and the inversion air. The inversion air has a slightly lower moist potential temperature than the layer beneath, since cooler mixtures are needed if the entrainment process is to proceed, but they will have a higher absolute potential temperature. As the mixed parcels descend, the depleted liquid-water content will evaporate and the parcel will come to rest once the water is gone. Further entrainment from the surrounding cloud may allow the descent to continue. The final mixture will always consist of a mixture of the same two components of air, so the moist potential temperature of the resulting dry parcel will always be about equal to the value for the undiluted cloud itself, since the inversion air is only a little lower in value. However, the absolute potential temperature of the inversion air is higher than in the cloud, so the absolute potential temperature of the final mixtures will be higher at levels where the higher liquid-water content requires more inversion air to evaporate the water. The absolute potential temperature will increase up through the region formerly filled with cloud, while the moist potential temperature will be about constant or slightly decrease. Below the former cloud base the air will initially have approximately the surface properties.

When new cloud is forming beneath the stratus cloud layer, the same process continues with some of the new cloud, while some continues to rise up into the stratus layer. The return flow ensures that the temperature structure produced in this way subsides towards the sea surface so that the absolute potential temperature will become stable at all levels. Thus, condensing parcels carry moisture up to various heights and
change the environmental structure towards a constant moist potential temperature equal to that of the source of the rising parcels. The net result is that the rising condensing parcels do not change the temperature stability of the layer but continue to carry up more water.

6. CONDITIONS ABOVE THE CLOUD

(a) Cloud-top radiative temperatures

The cloud-top radiative temperature measured from the aircraft varies much less than the sea surface temperature, with a slight increase over the warmer surface between soundings B and C, but little change over the warmer water from D to E. The cloud-top height changes little, being slightly lower for C and higher for D, but whether the slight decrease in cloud-top temperature is due to the greater height is difficult to determine.

(b) The air above the clouds

The prominent notches in moist potential temperature above cloud are an interesting feature, which gives additional information. As the satellite photographs show, a higher layer of stratus cloud is moving north-east across the lower marine boundary layer and has just cleared the area. When the soundings were made these showed no cloud water at those levels so the area was then clear. However, sounding A was in air just clear of the trailing edge of the upper cloud layer, in an area where entrainment processes were likely to have been active in evaporating the upper cloud edge.

Figure 6 shows the soundings, with the absolute potential temperature, the moist potential temperature, the mixing ratio and the wind vectors plotted side by side so that the variations can be related. Both the upward and the downward soundings at each position are shown for soundings A through D. It can now be seen that for sounding A the sharp dips in moist potential temperature above cloud are mostly associated with dips in water-vapour mixing ratio.

Such depleted notches in the water-vapour plot tend to fill in with time as shown in soundings B, C and D, which are successively further from the upper cloud edge. Thus an active process is needed to produce them near the cloud edge. Now the only source of drier air is from higher up in the atmosphere and the mixing ratio cannot be changed by radiative heating or cooling, so the drier air parcels near the current cloud top must have been brought down from above the second cloud layer as it evaporated. As discussed earlier, drier air mixed in from just above the cloud tends to travel through the cloud and accumulate at its lower surface, so this may provide an explanation. While this process could give the sharp decreases in mixing ratio, it would also leave air with the moist potential temperature of the drier air from higher up, and for the sounding A-up this is close to what actually happens. This structure is therefore likely the result of the evaporation by entrainment of the earlier overlying cloud layer.

In the case of A-down, however, the moist potential temperature appears to be slightly less than anywhere above. Since the notches lie on the warmer edge of a rapidly decreasing clear-air layer, this is the condition where thermal radiative transfer may produce cooling, and hence the temperature decrease.

There is sometimes a small increase in moist potential temperature in the clear air right against the cloud tops and then a decrease where the warmer inversion air finishes its steepest rise in the inversion. If radiative transfer cools the inversion air because of the much cooler cloud tops just below, and, at the same time, the cooler air below against the cloud top heats, this could provide an explanation for which no alternative is immediately apparent.
The spike in vapour mixing ratio, which is seen in several soundings at cloud top, is probably also the result of radiative heating as discussed above. The accompanying rise in moist potential temperature to higher values than elsewhere in the soundings seems to confirm this. The spikes in mixing ratio seen at the cloud tops are from the increase in mixing ratio resulting from the higher temperature of the saturated air in the top few metres of the cloud and in the few metres in the clear air just above.

7. CLOUD-TOP TEMPERATURES

Now let us look in more detail at the actual temperature of the cloud tops. Figure 7 shows the temperature and cloud liquid-water content on an expanded height scale and at ten samples per second. The temperature has been normalized to the pressure at 1000 m to remove pressure lapse-rate effects. The interesting result is that, with clear air above the clouds, the temperature, in all but one case, increases in the cloud towards cloud top, by more than 2 degC in some cases. In B-down the warmer layer extends about 50 m down into the cloud, while for B-up the temperature plot is constant until it stops at the level where the cloud liquid water becomes zero. In B the entrainment rate is very low as the adiabatic liquid-water profile in Fig. 5 shows. A recent intermittent episode of entrainment may have just renewed the cloud top in B-up and the warming effects may not have had time to take hold. It is noteworthy that sounding B-up is only a mile or so from B-down. All the processes associated with these clouds appear to operate intermittently, while driven towards equilibrium by the mixing processes discussed above or the radiative heating of the air just above cloud top.

These data contradict the widely maintained assertion that the cloud tops are cooler than the cloud below. The radiative transfer, which seems to be the only plausible
Figure 7. Temperatures and liquid-water mixing ratios are displayed on expanded vertical scales and with data at ten samples per second. The temperature has been converted to the reference pressure at 1000 m altitude, to remove changes due to pressure alone, and temperature is displayed only where there is cloud water. Except for sounding B-up, the top cloud layers where there is liquid water have a higher temperature than the cloud below. The dashed horizontal lines give the 1000 m altitude levels, and 100 m above and below are shown, except for D-down where the cloud top reaches 1100 m so the range is plotted from 800 m to 1200 m. Scales are the same everywhere except for the altitude in D-down where, because the cloud top was higher, the scale was changed to keep the 1000 m line at the same level as in the other graphs. B-up appears to be freshly formed on top from a recent entrainment event and has not yet adapted to the usual warmer cloud-top temperatures.

explanation of these effects, may well be due to the active contributions of very high-absorption lines when the temperature gradients are very large.

8. SUBSTRATUS CLOUDY CONDENSATION

The observed patches of cloud below the marine stratus layer were mentioned above in discussing soundings B and D. Figure 8 shows some details of the subcloud condensation mentioned in the discussion of Fig. 5. In the sounding D-up taken just after 7 h 30 min local time there is a region of liquid water below the base of the regular cloud layer. The essential role of these buoyant saturated blobs in building cloud down towards the surface has been discussed earlier. The liquid-water mixing ratio is plotted in time sequence using the aircraft forward air speed set at 85 m s⁻¹ as the abscissa, so the approximate width of such features can be seen. The lowest plot in the figure shows the liquid-water mixing ratio, starting at time 7 h 51 min 22 s at the altitude where water was first encountered after beginning the sounding climb up from 170 m altitude. This region of subcloud condensation was encountered earlier around 5 h on the first path to the north-west below cloud. It was then clearly visible in the aircraft running lights.

By plotting these data measured by the Forward Scattering Spectrometer Probe (FSSP) cloud-droplet probe against the aircraft path distance, the areas with significant numbers of drops can be seen to be in small regions a few tens to hundreds of metres across. After encountering the first liquid water near the start of the graph, the regular cloud base begins at 5000 m on the abscissa. The maximum liquid-water mixing ratio in the parcels below cloud increases to near 3000 m on the abscissa, with roughly the slope of the adiabatic liquid water which has been superimposed as a line on the graph from
Figure 8. Cloud-droplet average diameters, concentrations and liquid-water mixing ratios (left-hand axis scale) from the Forward Scattering Spectrometer Probe (FSSP), which measures drops less than about 40 μm in diameter. The averaged sample diameters from the one-dimensional probe are also shown. These drops are greater than about 50 μm in diameter and exclude the drops seen by the FSSP. These data are for sounding D-up, and cover the subcloud condensation shown in Fig. 5. The water is not actually in a single block as the lower sampling rate showed in Fig. 5, but is in isolated blobs less than about 200 m across. The main cloud base is beyond 5000 m and shows the substantially smaller increase of liquid-water content with height as compared to a well-mixed adiabatic layer. In the subcloud condensation up to 4000 m (on the abscissa), the peaks in liquid water appear to follow an adiabatic increase. The cloud-drop sizes increase upward in the main cloud, whereas the large drops decrease in size with height. These drops produce drizzle and are biggest in the saturated blocks below cloud.

near the regular cloud base at 5000 m and above. Thus the subcloud blobs of condensed drops are probably rising upwards with small regions remaining roughly adiabatic, but with strong entrainment of subsaturated air from their surroundings greatly diluting most parcels.

The average cloud-drop diameter is calculated for each FSSP sample and increases in the cloud from about 5 μm near cloud base to 15–20 μm near cloud top. This is to be expected from the almost adiabatic water release in the less diluted parcels as their height above cloud base increases. In the condensation below the cloud base the drop diameters increase from about 8 to 12 μm. The drop number concentrations have maximum drop numbers of about 200 drops cm⁻³, with a severe reduction in many parcels characteristic of entrainment dilution.

The other interesting feature is the diameters of the larger drops as measured by the 1-D probe, which largely ignores the cloud drops measured by the FSSP but measures the very low concentrations of drops greater than about 50 μm, at about 100 drops l⁻¹ or so. The average diameter of such drops is highest in the saturated regions below cloud and is 100 μm or more. These drops decrease in size towards 50 μm near cloud top, which is just about big enough for coalescence to start rapid growth. Such drops can start drizzle, or rain if the cloud is deep enough.
9. DISCUSSION

The dynamical development of the stratus cloud layer, as it flows over areas of ocean where the sea surface temperature gives alternate patches of cooler and warmer water, seems to be readily explained on the basis of simple physics once a few new processes are recognized.

(a) Vertical transport in and below the stratus cloud layer

The evidence presented here demonstrates that moisture can be carried up through atmospheric layers below cloud which are stable, on average, to dry convection, as reported by Telford (1992). This occurs without appreciably changing the stable temperature structure of the layers, which become moister. This evidence makes it clear that positive potential-temperature gradients do not prevent upward fluxes of horizontal momentum, and that moisture transport continues in such stable air. The condensed clouds carrying this moisture are only transitory (maybe sometimes only 'misty' patches) as they proceed intermittently up into the air above. Substantial clouds below the stratus layer have been reported previously, but the observations presented here suggest that such transport is always active in some form, except over warmer water. It seems possible that condensation produces latent heating sufficiently effective in lifting moist parcels, even though they continue to mix in increasingly potentially warmer, subsaturated surrounding air. This evaporates the drops and keeps the liquid-water mixing ratio at quite small values as the parcels rise into the potentially warmer surroundings.

Gradient diffusion based on an average structure does not appear to be a sufficiently adequate representation of moisture exchange processes since coherent isolated patches of moisture seem to play the essential role. Although turbulent eddy fluxes have been of use in description of the surface boundary layer, we note that, even in completely dry convection above about 50 m over land, the plume structure generates counter-gradient fluxes. For example, heat rises against the greater average potential temperature higher up (Warner and Telford 1967). This is the result of the vertical heat transfer in isolated plumes. In the case of marine stratus the intermittent nature of cloud-top entrainment and the intermittent formation of moisture-laden surface blobs make it essential to observe the changes in the boundary layer over time in order to integrate out the myriad of random excursions as the fluid adjusts. While technically a flux can be defined in any changing fluid, it seems that unless it is determined as a direct result of a physical process not originating in the field of fluid itself, it is not easy to apply. It is becoming clear that turbulent processes adapt to respond to the local instabilities from point to point and from moment to moment, which in turn are the result of the intermittent changes resulting from the fluxes themselves. Thus, while the changes in the fluid layers are determined by the adaption of the layer to the boundary conditions, representative fluxes do not follow as a result of the conditions in the fluid. The instantaneous fluxes can be highly variable, and, as their effect spreads through the fluid, this can modify later fluxes, and so the overall conditions tend to change slowly and erratically from one state to the next, towards equilibrium with the boundary conditions.

Boundary-layer data over land, taken at intervals through the morning hours, were analysed previously by Telford (1992). It was found that the activity of moisture-bearing buoyant elements carried moisture from the surface layers to well above the level to which the heat was rising. The temperature rises slowly in the stable overlying air (usually present in dry weather after overnight conditioning) to a level which has the same potential temperature as the surface, remaining roughly uniform with height up to that level. However, above that level, in the stable air, there is a dramatic increase in
vapour mixing ratio. The first clouds begin to form well above the level to which heat is penetrating, and even in cases where there is no cloud formation the moisture rises until a lapse rate is encountered which is more stable than the wet adiabat. At this level, moisture can accumulate from these buoyant parcels until clouds eventually form.

The data presented here were obtained from observations planned to follow the development with time of stratus cloud by sampling the cloud layer at various places after it has passed over areas of sea water at different temperatures. As is discussed above, the upward transport of moisture is by parcels from the surface moving up through the stable layers above, sometimes well into the cloud. Such a mechanism is needed to explain the observations, together with diluted parcels descending from cloud top, and thermal radiative exchange in clear air above. This mechanism offers an explanation for the lowering of the cloud base over cooler water (reported before by Rogers and Telford 1986). It also explains the dip in absolute potential temperature near the centre of the cloud layer when the surface is warm, and when the constant moist potential temperature extends from near the surface to cloud top.

As the subcloud condensation continues to accumulate in any one region, the cloud base will lower as the added liquid water extends closer to the surface. The temperature and buoyancy at the top of the subcloud condensation will increase as its liquid-water mixing ratio increases, until it is buoyant enough to penetrate into the overlying stratus. If it is buoyant at stratus base it will usually be buoyant right through the stratus cloud layer and will overshoot through the inversion above. If the sea surface were too cold to allow the subcloud parcels to rise up to the stratus, the whole substratus layer would eventually form another cloud layer, or fill with fog. It is far from clear how anything except the accumulated effects of sea surface temperature and the temperature and moisture in the air above the inversion can influence the stratus cloud layer. Solar heating had little influence on these data, and similar behaviour has been observed around midday. Conditions controlling cloud-top height and thickness were discussed by Telford and Chai (1984).

(b) Cloud-top temperatures

Any understanding of these clouds needs to include the observation herein reported that the tops of the clouds are warmer than the cloud below both at night and in the early morning, when the radiation from the sun has little influence. These observations are clearly at variance with the widely claimed assertion that thermal radiation cools the cloud tops and hence induces downward convection over cooler water. The probable explanation is that thermal radiative transfer involves some very intense absorption lines which are effective over short distances only when the temperature gradient is very high, as occurs at stratus cloud tops (Siems et al. 1993).

A marked increase in sea surface temperature promotes surface thermal convection and rising warmer in-cloud parcels which reach the top interface where they greatly increase the intake of dry air from above, both by better penetration, because they are warmer, and probably by turbulent stirring. The enhanced entrainment may convert the stratus cloud layer into cumulus clouds, as was observed in an offshore fog blowing west from the Californian coast (Telford and Chai 1993b). These freshly formed cumulus clouds soon totally evaporated because of the continued entrainment of subsaturated air from above. Wang and Albrecht (1994) have also studied these cloud-top entrainment processes. They concluded that, to determine cloud-top entrainment, information about the whole boundary layer is needed, not just a specification of the cloud-top interface. This paper strongly supports that conclusion, and further, suggests that the entrainment process is intermittent, that radiative cooling of the clear air above the cloud seems
to be involved and that cloud-top wind shear has little influence. Discussions of the dissipation of marine stratus cloud layers assume that no consideration need be given to the sea surface temperature or the vertical temperature and moisture structure of the overlying air. This work demonstrates that these factors are important and may well actually determine the overall outcome.

10. Conclusion

These field observations of marine stratus clouds were specifically designed to test the hypothesis that cloud-top thermal radiative cooling plays a role in the cloud formation and maintenance. While some previous reported measurements seemed to be in conflict with this general thesis, these direct observations demonstrate that other approaches need to be explored. In particular, thermal radiative transfer in the clear air across steep temperature changes needs further investigation. The process involving moisture transport upward from the surface was observed directly in the form of actual cloudy parcels leading to an explanation of the downward growth of the base of stratus clouds after moving over cooler water. The saturated parcels which transport moisture from the sea surface against a stable vertical temperature gradient were, indeed, found carrying drops upwards below cloud base. The accumulation of stratus saturated regions containing drops appears to result from the sea surface being cool enough that some of the rising parcels from the surface have insufficient buoyancy to reach the cloud despite the support from latent-heat release.

The top layers of cloud were found to be warmer than the cloud below in most cases (none cooler), so downward convection for this reason could not be active. Furthermore, as has been seen previously, the liquid-water mixing ratio in cloud growing downwards over cooler water was not adiabatic, but was higher lower down. The cloud was therefore not well mixed by convective overturning. This happened when the rising saturated parcels were cool enough to accumulate near or below the stratus cloud base. Above the stratus clouds in the clear air at the far northern end of the sampling region, the unusual vertical temperature and moisture structure can likely be explained by entrainment as the earlier higher cloud layer evaporated, as subsaturated air from further above mixed in.

Immediately above the cloud tops, thermal radiative transfer between warm and cooler clear cloud-free air, where the temperature gradient is very large, may be the cause of the increase in moist potential temperature usually found there which cannot be explained by mixing. Improved thermal radiative-transfer models with very high vertical resolution seem to be of high priority as does measurement of local gradients of flux divergence.

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