E-ε turbulence closure study of diurnal variation in a stratocumulus-capped marine boundary layer

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ABSTRACT

A model simulation is presented of the diurnal cycle of a marine stratocumulus-capped boundary layer. The model results are compared with observations obtained during the 1987 FIRE marine stratocumulus project, made from San Nicolas island, off the coast of California. Both, the simulation and observations show a marked diurnal variation in cloud properties, as a result of the decoupling of the boundary layer into a separate cloud and a sub-cloud layer. In the model simulation the decoupling is caused by the absorption of the solar radiation in the cloud layer.

1 INTRODUCTION

The important physical mechanisms which control the structure and type of stratocumulus are: cloud top entrainment instability (CTEI), diurnal decoupling and subsequent clearing due to solar absorption and mixing with dry air from above the inversion, drizzle, mesoscale circulations and subsidence.

We will focus on the effect of solar absorption on the decoupling and the effect of subsidence. Solar heating in the cloud can cause decoupling of the cloud layer from the sub-cloud layer. In a given situation, the tendency of the cloud layer and sub cloud layer to become decoupled is promoted by significant shortwave absorption by the cloud and by small surface buoyancy fluxes. This implies a seasonal dependence, decoupling being more likely in summer than in winter at mid-latitudes (Duynkerke, 1990). However, the sun may remain sufficiently high in the sub tropics for decoupling to remain possible at any time of the year.

During FIRE (First ISCCP (International Satellite Cloud Climatology Project) Regional Experiment) this decoupling was manifest as a weak stable layer that was observed below the main stratocumulus deck (Betts,1989; Hignett, 1991). Below this stable layer were very often cumuli that penetrated through this stable layer into the stratocumulus above. This can under certain conditions lead to the breakup of the solid stratocumulus deck. Turbulence closure models (Bougeault,1985; Duynkerke and Driedonks,1987; Duynkerke, 1990) predict this decoupling and simulate the thinning of the cloud layer during the afternoon. Here we will assess in some more detail the dynamical role.

If the subsidence is sufficiently strong, the inversion height can be lower than the mixed-layer lifting condensation level (LCL). In FIRE I the large-scale vertical velocity field was not measured systematically. This makes direct comparison of model results and observations difficult.

The purpose of this paper is to study the diurnal cycle of a marine stratocumulus layer, and make a comparison between the results of a model simulation (Duynkerke and Driedonks, 1987, 1988; Duynkerke, 1988) and detailed observations. We will use the observations made during the 1987 FIRE marine stratocumulus experiment. More specifically, we will use: cloud base height, inversion height, liquid water path, soundings of temperature and humidity, etc. Moreover, we will employ turbulence measurements made using instruments attached to the cable of a tethered balloon (Hignett, 1991). Some of the turbulence data was already discussed by Hignett (1991), part of the data is presented here for the first time.

2 MODEL DESCRIPTION

The model has been described in detail in Duynkerke and Driedonks (1987, 1988) and Duynkerke (1988). The one-dimensional model uses ensemble-averaged equations for the horizontal velocities (u and v), the wet equivalent potential temperature (θ_q, see Pointin 1984) and the total water content (q_w). The vertical velocity must be prescribed. The turbulent fluxes are modeled with the gradient approach, in which the exchange coefficient is calculated from the turbulent kinetic energy (E) and the viscous dissipation (ε), the so-called E-ε model (Duynkerke, 1988). In the entropy equation the heating or cooling due to the radiative flux divergence is included; a model is used for both longwave and shortwave radiation.

For longwave radiation we use the emissivity or "grey-body" approximation to calculate the longwave radiative flux. The effect of water vapor, carbon dioxide and liquid water on the emissivity is taken into account. The shortwave model includes Rayleigh scattering, absorption by atmospheric gases (water vapor, ozone and CO_2) and absorption and scattering by cloud droplets. Further details on the radiation model are given in Duynkerke and Driedonks (1988).

3 INITIALIZATION

The model simulation was started at 1000 UTC 14 July 1987 just before the observations of Hignett (1991). The location of San Nicolas island is 33.25 N and 119.5 W (Figure 1). At the sea surface the roughness length is set to 2 x 10^{-4} m and the albedo to 0.05. The sea surface temperature is taken as 288.4 K which is obtained from extrapolating the observed sensible heat fluxes with the flux-profile relationships towards the surface. The specific humidity at the sea surface is set to its saturated value at the sea surface temperature.

In the one-dimensional model the subsidence has to be prescribed. Observations of the large scale velocity field or vertical velocity field are lacking. Neiburger (1960) showed that climatologically the normal sea level pressure distribution over the eastern part of the North Pacific Ocean for July is anticyclonic with the pressure centre at 38° N and 150° W. Moreover, the winds deviate far from the isobars over this area, giving an area of significant divergence off the California coast. While the divergence has a maximum along the California coast it decreases in all directions over the ocean, but the divergence remains positive over practically all the eastern part of the North Pacific

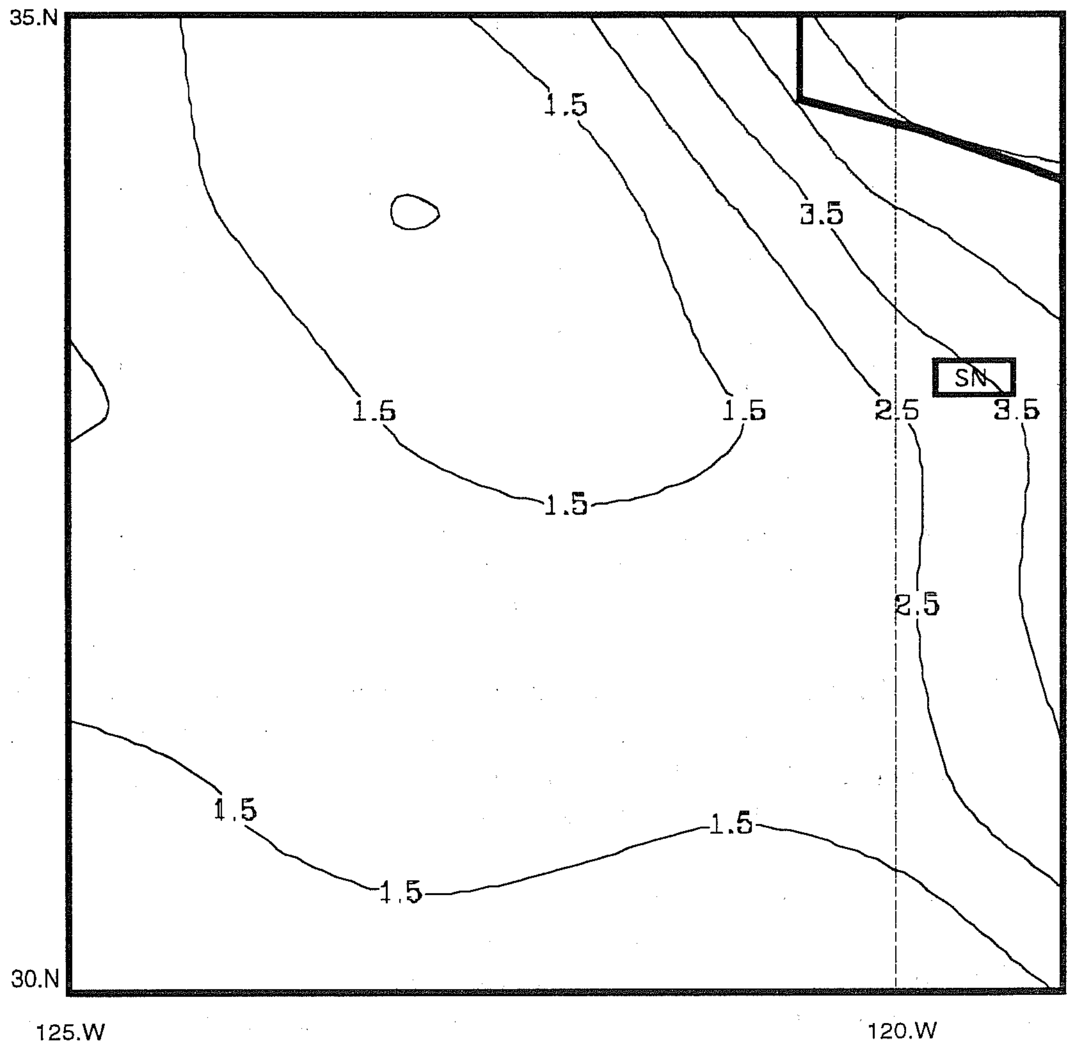


Figure 1 The time-averaged divergence field (x 10^6 s^{-1}) for the period 1 to 19 July 1987 calculated from the ECMWF initialised analysis. Indicated is also the location of San Nicolas island (SN).

Ocean. From Figure 6 of Neiburger (1960) we get an average value for the divergence of 6×10^{-6} s^{-1} for July over San Nicolas island.

Therefore we analysed the subsidence from the ECMWF model in the FIRE area from 119° W to 125° W and from 30° N to 35° N. From the ECMWF model we have analysis fields every 6 hours for the period from 1 to 19 July 1987. The averaged divergence field over this period is shown in Figure 1. This is similar to the results found by Neiburger (1960) who found for July divergences up 6×10^{-6} s^{-1} off the coast of California and decreasing divergences further east over the Pacific. In order to study the diurnal variation in the divergence at San Nicolas island the 6-hourly values of the divergence at 119°W 33°N are shown as small dots in Figure 2. The 19 day-averages at a

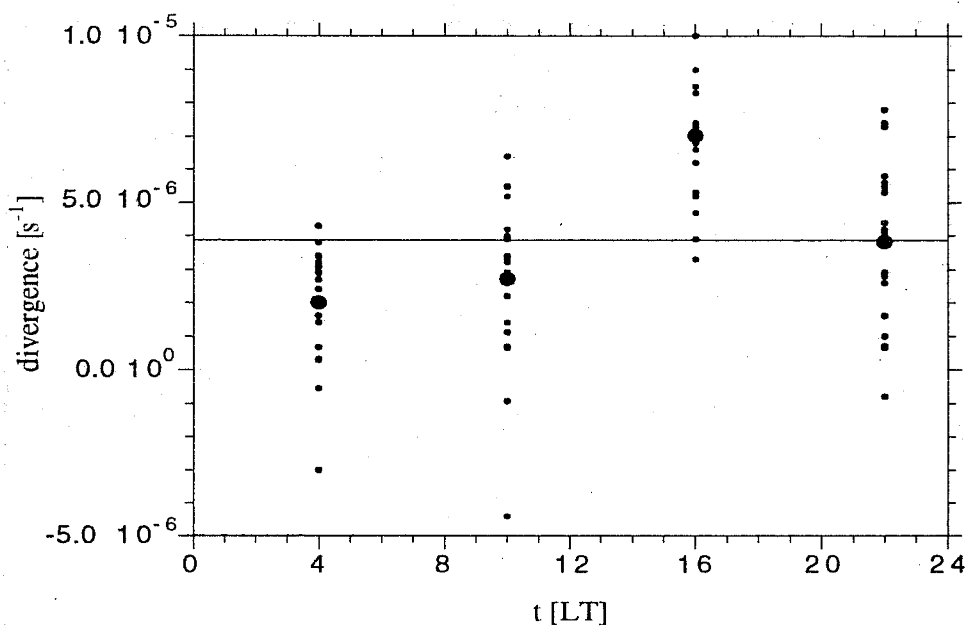


Figure 2 The 6-hourly values of the divergence from the ECMWF model at 33° N 119°W for the period 1 to 19 July 1987 (small dots), the averge over 19 days at a certain hour are shown as large dots and the average over the complete period (3.9×10^{-6} s^{-1}) is shown as a horizontal line.

certain hour are shown as large dots. Whereas the divergence averaged over the whole period (3.9×10^{-6} s^{-1}) is shown as a horizontal line.

The averaged value of the divergence corresponds, at 600m height to a vertical velocity of -0.23 cm/s. On average this must be balanced by advection and entrainment (w_e)

$$\frac{\partial h}{\partial t} = - u \frac{\partial h}{\partial x} - v \frac{\partial h}{\partial y} + w_e - w , \qquad (1.1)$$

in which h is the inversion height, w the large scale vertical velocity and w_e is the entrainment velocity defined by:

$$w_e = - \overline{w'c'} / \Delta c, \qquad (1.2)$$

The diurnal variation in subsidence suggests that the inversion is descending from 1200 to 2200 LT as observed. However, the amplitude of the diurnal variation in divergence gives an amplitude of about 40 m whereas the observed amplitude in cloud-top height is about 100 m.

For the day under consideration the time-averaged divergence from the ECMWF model is 3.5×10^{-6} s^{-1}, this constant value was used throughout the model run. The observed mean wind direction in the boundary layer is about 290 degrees (Hignett, 1991). In the upstream direction the divergence from the ECMWF analysis is nearly constant over several hundreds of kilometres. So, if the simulation would be in a lagrangian framework, in which the column of air is advected along with the mean wind, the divergence would be fairly constant along the trajectory.The geostrophic wind speed was set to 5.5 m/s (Hignett, 1991) and the wind direction to 290 degrees. The wind is therefore blowing almost parallel to the isotherms of the sea surface temperature (Duynkerke and Hignett, 1993). This means that the surface temperature hardly changes as the air flows south-eastwards. Therefore it is quite realistic to keep the sea surface temperature in the one-dimensional model constant with time. The initial wet equivalent potential temperature and total specific humidity profile are taken to fit closely the observed profiles. We used the cross-chain Loran atmospheric sounding system (CLASS) data up to 2 km of Schubert et al. (1987a). The vertical profiles of θ_q and q_w at 0014 and 1158 UTC on 14 July and 0015, 1200, 1703 and 1935 UTC on 15 July are shown in Duynkerke and Hignett (1993). There is a clear well mixed boundary layer with a strong capping inversion. The boundary layer is moist with a dry overlying atmosphere. The temperature jump at the inversion is about 12 K (Hignett, 1991). For the initial model profiles we take θ_q = 314 K within the boundary layer and θ_q = 323 K above it. The specific humidity is taken as 9.6×10^{-3} kg/kg in the boundary layer and just above it is set to 7.6×10^{-3} kg/kg, with a lapse rate of -4×10^{-6} m^{-1} above the inversion. The model top is taken as 1722m.

4 RESULTS AND DISCUSSION

We will concentrate on the diurnal variation of the stratocumulus deck due to the decoupling of the cloud layer from the sub-cloud layer as observed during FIRE. More specifically we will use the observations of 14 and 15 July 1987 and compare those with a model simulation. During this period a steady north-westerly flow was observed at San Nicolas island. Which reduces the possibility of a coastal effect on the diurnal variation observed from the island.

In section 4a we will discuss the observed and simulated variation of cloud base, cloud top and liquid water path. In section 4b we will discuss in more detail the cause of the diurnal variation, that is the decoupling of the cloud layer from the sub-cloud layer. Moreover, the modelled and observed turbulent quantities will be compared. In section 4c we will discuss the influence of the decoupling on the surface fluxes and the entrainment velocity.

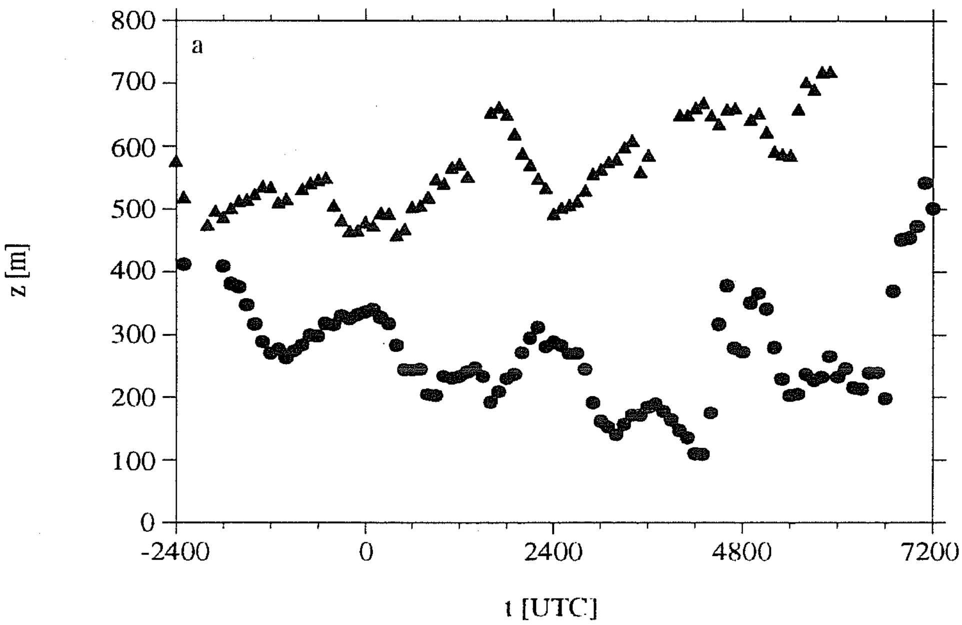
4.a DIURNAL VARIATION

In order to study the time evolution of clouds during the FIRE experiment several instruments were used on San Nicolas island to monitor cloud properties with a high temporal resolution. Cloud base height was measured by a Vaisala CT 12K laser ceilometer operated by Colorado State University as described by Schubert et al.(1987b). The ceilometer operated continuously and gave cloud base height every 30 seconds.

A sodar was used to estimate the inversion height (White, 1989). The height was determined where the maximum in the reflectivity of an acoustic signal occurred. This is at the height of the temperature inversion which is then associated with the height of cloud top.

The vertically integrated liquid water content was retrieved using a three-channel (20.6, 21.65 and 90.0 GHz) microwave radiometer as described by Hogg et al. (1983). A statistical algorithm converts recorded brightness temperature of the atmosphere into path-integrated liquid water (liquid water path). The absolute uncertainty in the instrument is estimated to be 20%. The radiometer yielded nearly continuous measurements throughout the FIRE period with a time resolution of 1 minute.

The variation of cloud base and top height, as retrieved from the ceilometer and the sodar, is shown in Figure 3a as a function of time for 13, 14, 15 and 16 July 1987. In order to indicate time continuously in the figures we have denoted the period from 0000 UTC 13 July to 2400 UTC 16 July as -2400 to 7200. We have used 0000 for 0000 UTC 14 July because we start our simulation on the 14 th of July. As can be seen from Figure 3a the cloud thickness increases during the period under consideration. Moreover there are variations in cloud thickness on the time scale of a day or less. We are mainly interested in the latter variations and not in the slow variations on the time scale of a day or more, which are probably due to changes in the synoptic circulation (subsidence) in the area under consideration. In order to highlight the diurnal variability of the cloud, rather than the slow changes due to the synoptic observations we have objectively removed the effect of the long term trend. We have detrended cloud top and base height by calculating for each hourly value the departure from the 24-h mean around that value. These departures were then added to the overall mean of the four-day period, producing the



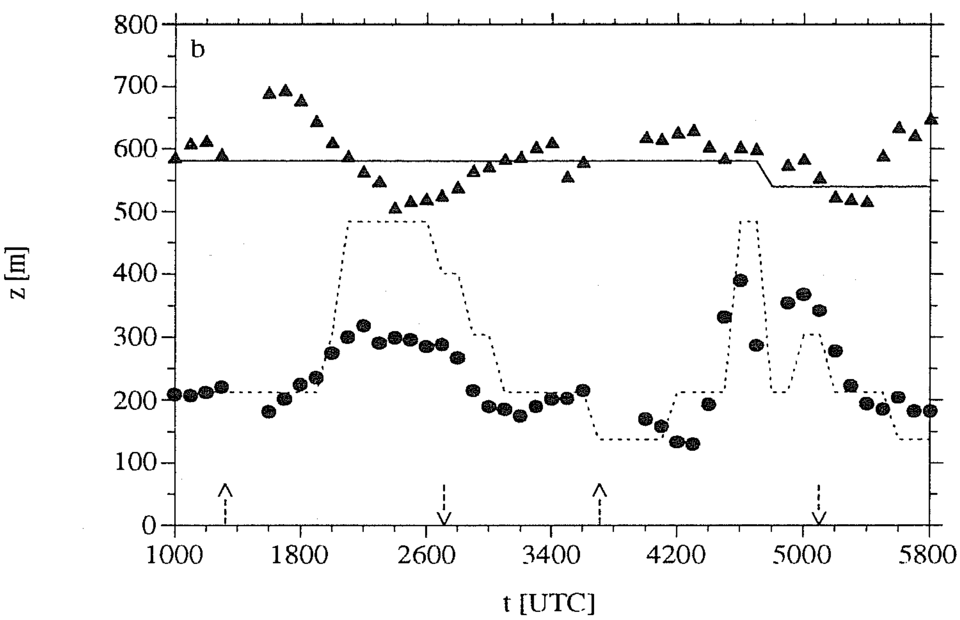


Figure 3 Time evolution of the observed cloud top and base height (a). The detrended observed (symbols) and simulated (lines) cloud top and base height as a function of time (b). Sunrise and sunset are indicated by an upward and downward arrow, respectively.

results shown in Figure 3b. The observed average cloud top height is 585 m and the average cloud base height is 238 m.

The simulated cloud base height (Figure 3b) show more or less the same variations as the observations with time. Noon is around 2000 UTC, with sunrise and sunset occurring at about 1300 and 2700 UTC, respectively. It is clear that both in the observations and model results the cloud base lifting sets in some hours after sunrise. The maximum cloud base lifting takes place in the late afternoon. After this the cloud base descent sets in lasting until midnight. It is clear that the diurnal changes do not follow the solar insolation directly, in the sense that they are not symmetrically around local solar noon. This is not surprising since the largest solar heating rates and therefore the largest changes in cloud base height are expected around solar noon. This can be seen in both the observations and simulation results shown in Figure 3b.

For the optical properties of the cloud the liquid water path (the vertically integrated liquid water content) is an important quantity to consider (Stephens, 1984). Typically the liquid water content varies linearly with height with the vertical gradient equal to $(\partial q_{sat}/\partial T)_T \partial T/\partial z$. This gives that the liquid water path is proportional to the cloud layer thickness squared (Blaskovic et al., 1991). The liquid water path will therefore show a much larger temporal variation than the cloud thickness. The observed and simulated liquid water path is shown in Figure 4 as a function of time. The simulated

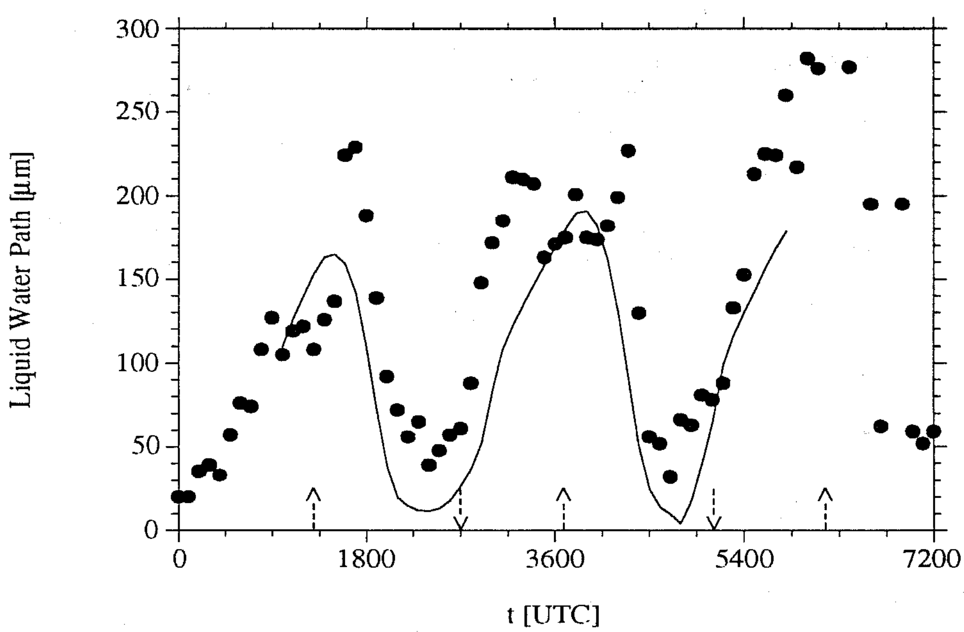


Figure 4 The observed (dots) hourly-averaged liquid water path compared with the model simulated (full line) liquid water path as a function of time.

liquid water path typically varies between 0.2 mm at around sunrise to about .02 mm some hours before sunset. The observations show a similar kind of diurnal variation. In the model simulation the liquid water path is typically very close to the adiabatic liquid water path, very similar to that in the observations (Albrecht et al., 1990). The largest difference between the simulated liquid water path and the adiabatic value occurs from 1500 to 2400 UTC, which is during the decoupling of the cloud layer from the sub-cloud layer.

4.b TURBULENT STRUCTURE AND VERTICAL PROFILES

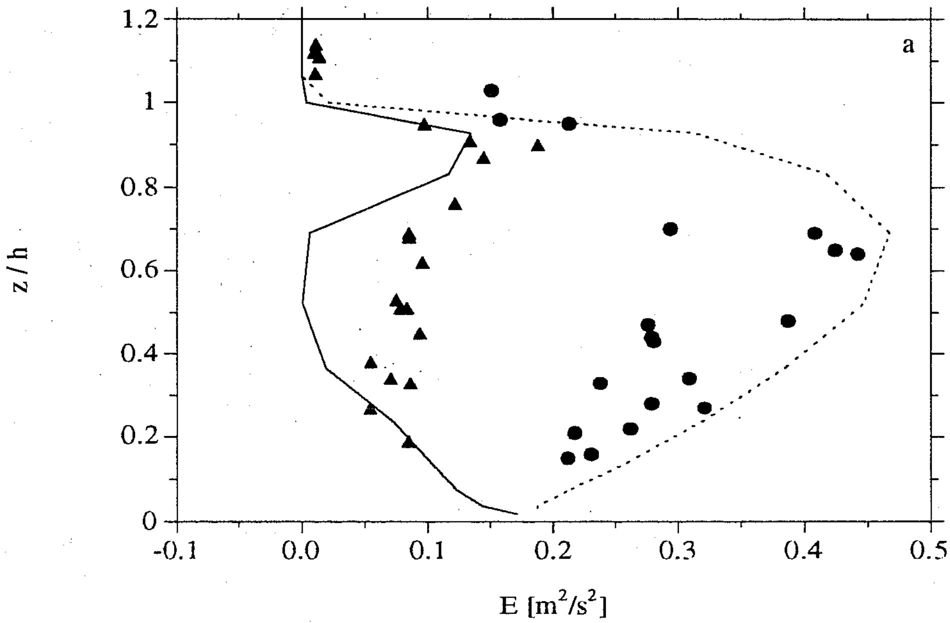
In the previous section we compared the simulated with the observed temporal evolution of cloud top height, cloud base height and liquid water path. In this section we will discuss in somewhat more detail what causes the diurnal variation. We will do this by looking at the simulated vertical structure of the boundary layer. Moreover, we make some comparisons with the observations of Hignett (1991).

Hignett (1991) presented measurements made during FIRE by instruments attached to the cable of a tethered balloon. The data were selected for a day and a night period, corresponding to 1830-2100 (14 July) for the day-time and 0530-0800 (15 July) for the night. Local noon occurred at about 2000. He showed that during day-time the boundary layer was decoupled into a separate cloudy layer and a surface (sub cloud) layer.

The periods chosen for comparison by Hignett (1991) were centred around local noon and midnight. These times are the extremes of the external forcing from solar radiation but do not correspond to the extremes of the resulting diurnal cycle of cloud thickness and liquid water path. However, these periods do represent decoupled and coupled cases and have been chosen for a direct comparison between the observations and simulation.

Hignett (1991) used the vertical velocity variance as an indicator of convective activity. Since in the model only the turbulent kinetic energy is calculated and we have no means to separate this into the variances in the different directions, we have plotted both the simulated and observed (both at 3200 and 4500) turbulent kinetic energy in Figure 5a. Whereas, the observed vertical velocity variance at these times is plotted in Figure 5b. It can be clearly seen that during the night the boundary layer is turbulent up to cloud top, whereas, during day-time the turbulent kinetic energy decreases from the surface upward to near zero below cloud base. The cloud layer is turbulent due to the convection initiated by the longwave radiative cooling at cloud top. The nocturnal boundary layer is thus a well mixed layer from the inversion to the surface, driven from cloud top (by longwave radiative cooling) in a manner analogous to that of a convective boundary layer heated from below.

All these features are also reflected in the turbulent fluxes of conserved quantities (q_w, θ_q), total water content and wet equivalent potential temperature (Pointin, 1984),



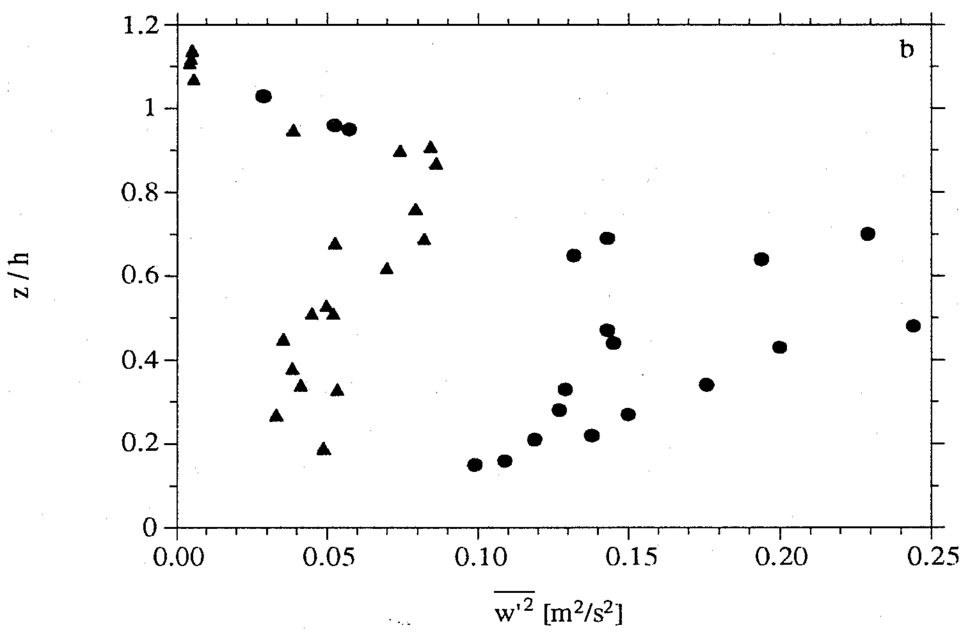


Figure 5 The simulated (lines) and observed (symbols) turbulent kinetic energy (a) and observed vertical velocity variance (b) at 3200 (dashed line and dots) and 4500 UTC (full line and triangles) as a function of dimensionless height.

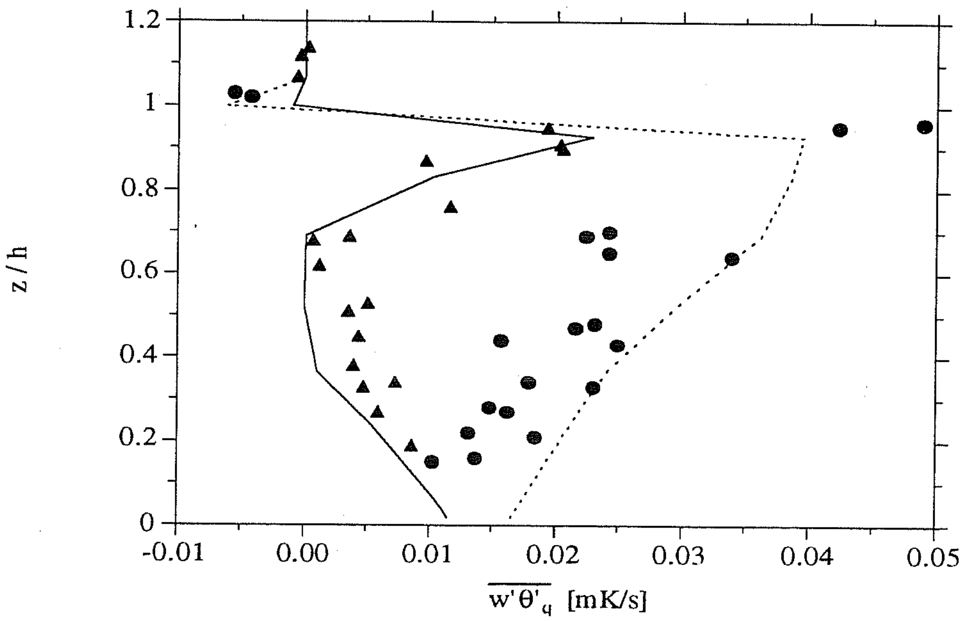


Figure 6 The simulated (lines) and observed (symbols) equivalent potential temperature flux at 3200 (dashed line and dots) and 4500 UTC (full line and triangles) as a function of dimensionless height.

respectively. As an example we have shown in Figure 6 the simulated and observed vertical profiles of the wet equivalent potential temperature flux. The simulated fluxes at the surface are positive. The flux maxima are located at cloud top, with the larger fluxes occurring during the night. The nocturnal data again give the appearance of a single well mixed layer driven by the longwave radiative cooling at cloud top. The day-time profile shows clearly two turbulent layers: a surface layer and a cloud layer.

The turbulent structure of the stratocumulus-topped atmospheric boundary layer can be studied in more detail through the inspection of the turbulent kinetic energy budget. In one-dimensional form the equation for the turbulent kinetic energy (Duynkerke and Driedonks, 1988) reads

$$\frac{\partial E}{\partial t} = \underbrace{-\overline{u'w'}\frac{\partial u}{\partial z} - \overline{v'w'}\frac{\partial v}{\partial z}}_{S} \underbrace{-g\frac{\overline{w'\rho'}}{\rho_0}}_{B} -w\frac{\partial E}{\partial z} \underbrace{-\frac{\partial}{\partial z}\left(\overline{w'E'} + \frac{\overline{w'p'}}{\rho_0}\right)}_{T} \underbrace{-\varepsilon}_{D}, \tag{4.1}$$

in which the terms on the right-hand side are the shear production (S), buoyancy production (B), vertical advection (subsidence), turbulent transport (T) and viscous dissipation (D).

In Figure 7a and b the simulated and observed terms in the turbulent kinetic energy budget are compared. For the observations the terms in (4.1) were evaluated as described by Nicholls (1985). During the night (3200 UTC, Figure 7a) the turbulence is mainly driven by the longwave radiative cooling at cloud top. This gives convection throughout the whole boundary layer as can be seen from the positive values of the buoyancy production from cloud top down to the sea surface. Close to the surface the buoyancy production is small because there is only a small difference between the sea surface and the air just above it. In the model the transport term is negative (sink) in the cloud layer and positive at the inversion and in the sub cloud layer. The observations show a somewhat opposite trend. However, this might be due to the fact that it only includes the divergence of $\overline{w'E'}$ and not the term $\overline{w'p'}/\rho_0$. The latter term cannot be evaluated from the observations because the pressure fluctuations are not measured. The modeled and observed transport term (T) do not represent the same quantities. The observed viscous dissipation is in the sub cloud layer larger than the simulated viscous dissipation. As a result it is clear that in the observations there is quite a large imbalance between the sum of all the terms; overall the picture appears similar to that described by Nicholls (1989).

During day-time (4500 UTC, Figure 7b) the longwave radiative cooling is about the same as during night-time because the cloud is still optically thick for longwave radiation. The main difference is the shortwave radiative heating of the cloud layer. This heating is mainly distributed over a layer with a thickness of about 200 m near cloud top. As a result the shortwave radiative heating offsets part of the longwave radiative cooling in the upper 50 m of the cloud layer. Below this the shortwave radiative heating is dominant. If the boundary layer remains well mixed this means that the buoyancy production near cloud base will diminish and, if the divergence in the shortwave radiation is large enough, it can make the buoyancy flux near cloud base negative. This implies that the convection is suppressed and as a result the turbulent transport of momentum, heat and moisture will be reduced. Both the observations and simulations (figure 7b) show that the buoyancy production is mainly localized inside the cloud layer, where it is balanced by turbulent transport and viscous dissipation. The simulated turbulent kinetic energy budget clearly shows a three layer structure in the boundary layer: a turbulent layer connected to the surface mainly driven by shear, a convective cloud layer mainly driven by longwave radiative cooling at cloud top and in between a transition layer which is near neutral to slightly stable, where all the production terms are very small. For this day-time case the simulated results are in rather good agreement with the observations. Moreover, the observed terms in the turbulent kinetic energy budget are in quite close in balance, which gives good confidence in the quality of the data.

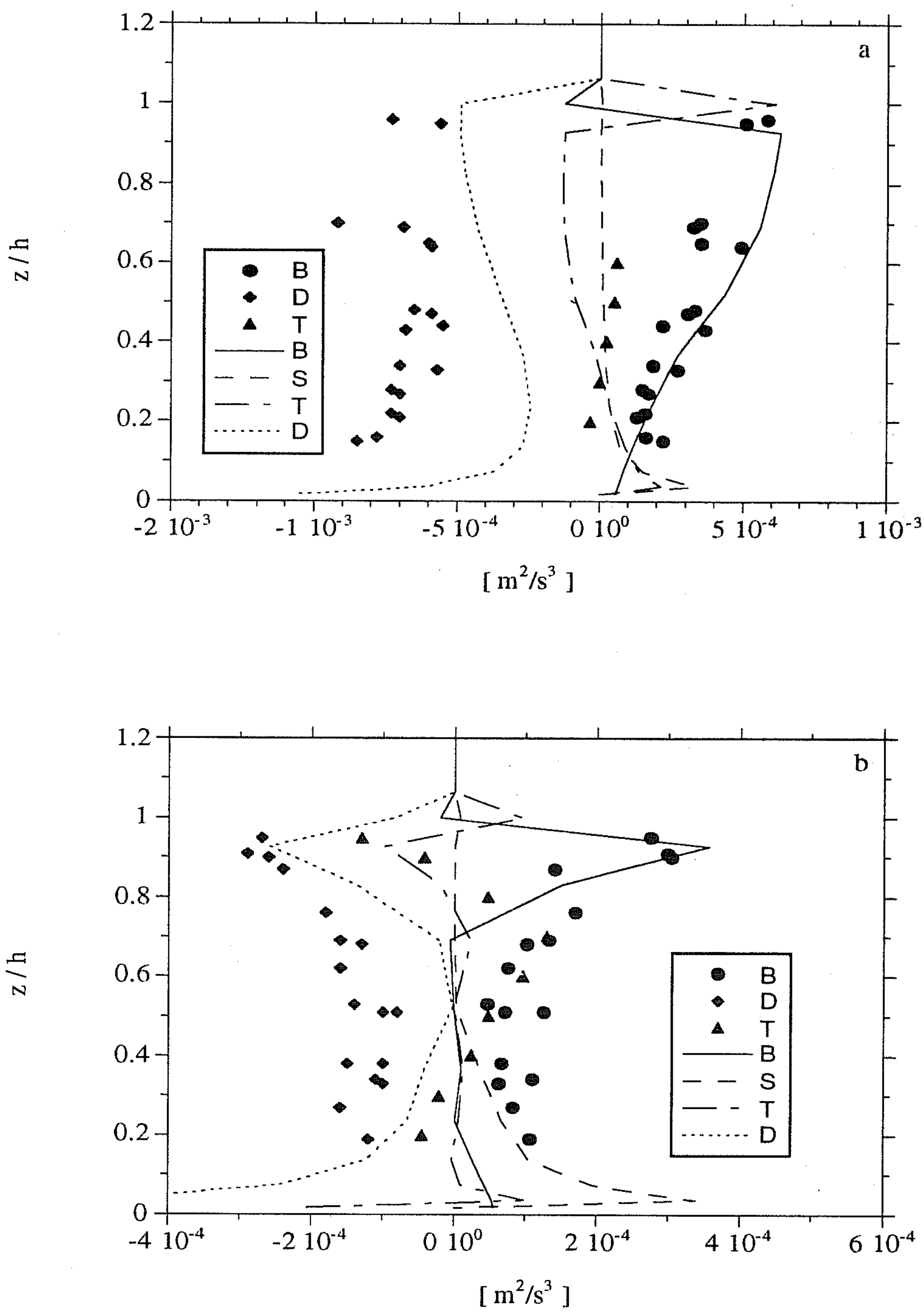


Figure 7 The simulated (lines) and observed (symbols) turbulent kinetic energy budget as a function of height (normalized by the inversion depth) for the night-time (a) and day-time (b) period: buoyancy production (B), shear production (S), turbulent transport (T) and viscous dissipation (D).

4.c CONSEQUENCES OF DIURNAL VARIATION

Both observations and detailed models are expected to provide guide-lines for developing simpler models. Standard mixed layer models are clearly inadequate to describe the decoupling of the cloud layer and the sub cloud layer. Turton and Nicholls (1987) developed a model in which the single mixed layer observed during night time was diagnosed to split into two mixed layers during day-time. The criterion used to diagnose this decoupling is that the maximum allowable energy loss against negative buoyancy is a fraction (40%) of the buoyancy produced in the cloud layer. In this way the height of the base of the mixed layer produced by the cloud convection can be calculated. Turton and Nicholls (1987) showed that the introduction of these two separately well mixed, but decoupled, layers produced a significant diurnal variation in cloud thickness. When separation is not allowed, the cloud thickness varies only slowly and the diurnal variation is mainly due to the variation in entrainment velocity.

In cloud-free mixed layers the entrainment velocity (w_e) is found to depend mainly upon the properties of the turbulence maintaining entrainment and the density difference across the interface ($\Delta\rho$). Many investigations have been carried out with turbulence being driven by a variety of processes: mechanical stirring with grids, surface shear stress, convection resulting from surface heating and by shear across the interface. Dimensional arguments (Turner, 1973) suggest that, in these experiments involving a

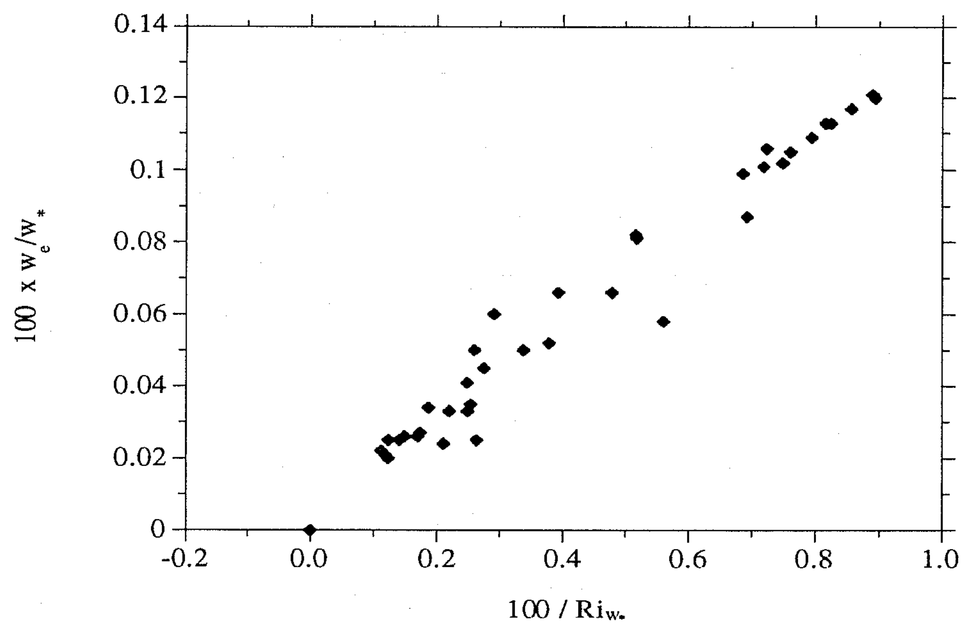


Figure 8 The simulated dimensionless entrainment velocity as a function of the inverse of the Richardson number given by (4.3).

dominant turbulence production mechanism with a length scale l and velocity scale U, the entrainment velocity scales as

$$w_e/U = f(Ri_U) = f(-\frac{g}{\rho_0}\frac{\Delta\rho\, l}{U^2}) \;, \qquad (4.2)$$

in which Ri_U is the Richardson number.

In geophysical applications l is usually taken as the mixed layer depth (h) and for the velocity scale u_* is used for shear-driven layers and w_* for a convectively driven layer. For the convectively driven case, which we will consider, the most widely used result is

$$w_e/w_* = a\, Ri_{w_*}^{-1} \;, \quad \text{with } Ri_{w_*} = \frac{g}{\theta_v}\frac{\Delta\theta_v\, h}{w_*^2} \;, \qquad (4.3)$$

with values of a =0.2 to 0.25 (Driedonks, 1982; Deardorff, 1983) for the cloud free case and model derived values of about 2 for the cloudy case (Nicholls and Turton, 1986). The hourly values from the model simulation are shown in Figure 8 and indicate a value of about 0.14 for a in (4.3). These values seem rather low compared with the values of

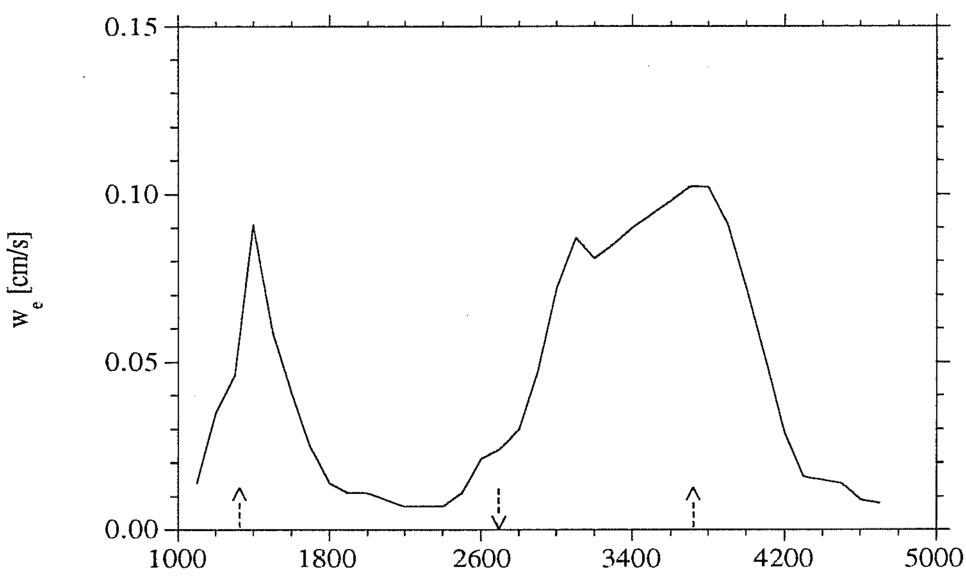


Figure 9 The entrainment velocity w_e in [cm/s] as a function of time.

Nicholls and Turton (1986). The variation of w_e as a function of time is shown in Figure 9. The entrainment velocity is thus very small during day-time and has a value of about 0.1 cm/s during night-time.

The diurnal variation in cloud and turbulence properties gives also a diurnal variation in the surface fluxes as shown in Figure 10. The sensible (H) and latent (LE)

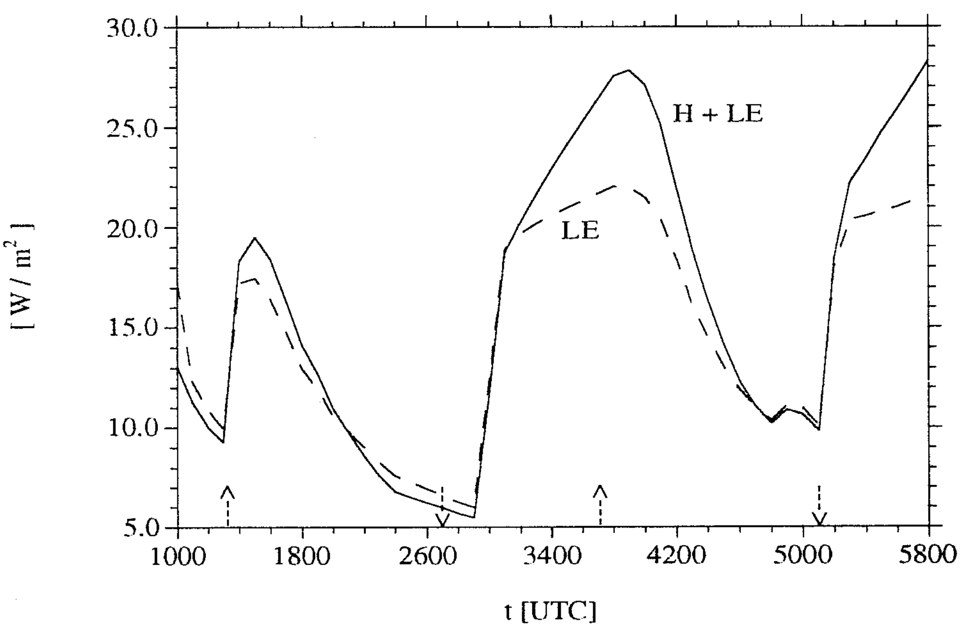


Figure 10 The latent heat flux LE (dashed line) and the sum of sensible and latent heat flux H+LE (full line) as a function of time.

heat flux show a maximum around sunrise and a minimum around sunset. This is mainly a result of the decoupling of the cloud layer from the sub cloud layer. During night-time the convection through the whole boundary layer is driven by the cloud top radiative cooling. This longwave cooling cools the boundary layer and as a result the temperature difference between the sea surface and the air just above it will increase giving rise to an increasing sensible heat flux with time. During day-time the cloud layer and the turbulent layer adjacent to the sea surface become decoupled, so that the longwave cooling can no longer cool the turbulent layer adjacent to the sea surface. Moreover, the dominant process which can change the temperature of this layer is the sensible heat flux from the sea surface. As a result the temperature of this layer increases and the temperature difference with the sea surface diminishes, with the result that the sensible heat flux decreases towards zero.

CONCLUSIONS

A comparison between the results of a model simulation and detailed measurements of the diurnal cycle of a marine stratocumulus-capped boundary layer during the 1987 FIRE have been made. In the model simulation the decoupling is caused by the absorption of solar radiation in the cloud layer. The decoupling of the boundary layer into a separate cloud and sub-cloud layer is most pronounced from noon to just before sunset. The decoupling is very similar to the previously reported observations of Nicholls (1984) and Nicholls and Leighton (1986).

Both, during day-time and night-time the simulated and observed turbulent kinetic energy budget are compared. The multiple-level turbulence measurements, made simultaneously from a series of probes attached to the cable of a tethered balloon, give a detailed vertical profile of the most important terms in the turbulent kinetic energy budget. Both, the observations and simulation reveal that during night-time the boundary layer is well-mixed. With a small surface buoyancy flux the turbulence is mainly driven by the longwave radiative cooling at cloud top, giving buoyancy production throughout the boundary layer.

During day-time the absorption of solar radiation in the cloud layer gives rise to the formation of a stable layer below cloud base. Both the and observations and simulation show that the buoyancy production is mainly localised in the cloud layer. The simulated turbulent kinetic energy budget clearly shows a three-layer structure in the boundary layer: a turbulent surface mixed layer driven by wind shear, a convective cloud layer and in between a transition layer. In this transition layer all production terms are very small and the stratification is from near neutral to slightly stable.

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