Evolution of the Subtropical Marine Boundary Layer: Comparison of Soundings over the Eastern Pacific from FIRE and HaRP

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ABSTRACT

The mean time rates of change of temperature, total water mixing ratio and ozone along airflow trajectories in the lower troposphere over the eastern Pacific are inferred by comparing aircraft soundings from the First ISCCP Regional Experiment (FIRE) and the Hawaiian Rainband Project (HaRP). Through the use of the estimated mean fluxes of temperature and total water mixing ratio, it is found that the tendency for stratus layers to grow or dissipate is very sensitive to the assumed turbulence structure below the capping inversion. A mixed-layer model that assumes a well-mixed boundary layer up to the capping inversion predicts a solid cloud layer extending all the way to Hawaii, whereas a model that allows decoupling predicts rapid dissipation of the stratus layer. It is concluded that stratus dissipation here is due to the slowdown of turbulent mixing throughout the layer below the capping inversion, not the drying out of a well-mixed layer; hence, the mixed-layer model cannot be expected to predict realistic cloud dissipation. The differences in ozone concentration observed in the boundary layer during HaRP and FIRE suggest a chemical loss of ozone of 3–8 ppb day$^{-1}$, corresponding to a lifetime of 3–9 days. This implies that ozone cannot be treated as a conserved tracer when dealing with ozone budgets over periods of days. The ozone sink is probably of photochemical origin, and it requires further investigation.

1. Introduction

Variations in the extent and distribution of marine stratiform clouds have a significant impact on the earth’s radiation budget. Thus, simple but accurate parameterizations of stratiform clouds are needed for realistic modeling of climate changes. Currently it is not clear to what extent the needs for simplicity and accuracy can be reconciled, since the processes involved are quite complex. The formation and dissipation of stratiform clouds depend on the interplay and balance among several processes: turbulent transport of heat and moisture from the sea surface, entrainment at the cloud top, radiative cooling/warming, precipitation, and subsidence. In recent years numerical models of varying degrees of complexity have been developed focusing on some of the above processes in detail, or dealing with most or all of them in some parameterized form (e.g., Bechtold et al. 1992; Bretherton 1992; Kogan et al. 1992; Krueger and McMurtry 1992; Moeng et al. 1992; Randall et al. 1992; Rogers and Koracin 1992; Siems et al. 1993; Wang et al. 1992; Yancho et al. 1992; MacVean and Mason 1990; Duyknerke 1989; Betts and Ridgway 1988; Turton and Nicholls 1987; Bougeault 1985).

In contrast to this abundance of numerical models, currently there are few observations that can be used to evaluate model predictions. Field experiments have usually been conducted in a fixed area, whereas boundary-layer models typically use a Lagrangian frame of reference. Lagrangian field experiments extending over several days are difficult to perform, particularly over the eastern Pacific where the air follows trajectories far removed from airports. Recently, attempts were made to carry out Lagrangian observations in the Atlantic Stratocumulus Transition Experiment (ASTEX) over the eastern Atlantic (30° to 40°N). Data from this field project are now being analyzed. The use of satellite data to deduce time evolution of marine stratus is promising (Pincus et al. 1992; Minnis et al. 1992), but the information derived this way is somewhat limited. Here we attempt another approach: namely, to infer the mean time rates of change in the marine boundary layer over the eastern Pacific from aircraft soundings from the First ISCCP (International Satellite Cloud Climatology Project) Regional Experiment (FIRE), conducted about 400 km off the California coast in the summer of 1987 (Albrecht et al. 1988), and from aircraft soundings 100–200 km upwind of the island of Hawaii.

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obtained during the Hawaiian Rain Project (HaRP) in the summer of 1990.

2. Airflow in the FIRE–HaRP area

We shall here assume that the air arriving in Hawaii during HaRP previously had properties similar to those observed during FIRE. This assumption is based on rather crude interpolations in space and time. Nevertheless, until data from more definitive Lagrangian experiments become available, the following sounding comparisons provide estimates of the mean changes occurring in the boundary layer as the air travels over the eastern Pacific Ocean for several days.

Observations indicate that the boundary-layer air arriving in Hawaii had previously traveled in the neighborhood of the FIRE research area. Figure 1 shows surface wind stress vectors and divergence over the Pacific Ocean in August averaged over the years 1966–1985 and during the year 1985 (from Striebert et al. 1992). The circles marked P and H indicate the FIRE and HaRP research areas. Monthly averages for July and August show that the airflow in the FIRE–HaRP area does not change much from year to year. It can be seen in Fig. 1 that the air that arrives in the HaRP area typically comes from a region about 6° west of the FIRE area, at about 130°W. The divergence along the airflow trajectory is $\sim 3 \times 10^{-6}$ s$^{-1}$, with an average value of $\sim 10^{-6}$ s$^{-1}$.

The typical aircraft research area during FIRE was about 400 km offshore, extending from 30°N to 37°N, 121°W to 126°W. (Flights near the coast and over San Nicholas Island are not included since airflow there follows different trajectories.) Aircraft soundings in the usual research area showed considerable variability, but there were no systematic changes in the east–west direction that could be attributed to persistent horizontal gradients in temperature, moisture, or the height of the capping inversion. Long-term averages (Neiburger et al. 1960; Riehl et al. 1951) suggest that the inversion should rise slowly toward the west, but in the FIRE research area, which covered only 5° in longitude, this effect was not significant compared to the day to day variability of the inversion height. This leads us to expect that the soundings 6° west of the FIRE research area should have a similar range of variability as the soundings in the FIRE area.

Observations from other field experiments in the FIRE and HaRP areas during summer suggest that the year to year variations in the summer soundings are also relatively minor. Soundings from the Dynamics and Chemistry of Marine Stratocumulus (DYCOMS) Experiment, which was conducted in the summer of 1985 in about the same area as FIRE, were not substantially different from those in FIRE. Similarly, soundings from the Hawaiian Warm Rain Project (HWRP), which was conducted in the summer of 1985 in the same area as HaRP, were usually within the range observed during HaRP, except for a few days when there was a hurricane nearby.

Thus, we presume that the differences in the datasets from FIRE and HaRP represent time rates of change within an air mass traveling along a Lagrangian trajectory, which passes 6° west of the FIRE research area and arrives in the HaRP research area. As defined by the average winds in Fig. 1, this trajectory is about 3000 km long. An air mass traveling at an average speed of 7 m s$^{-1}$ (as estimated from typical wind velocities during FIRE and HaRP and Fig. 1) should reach the HaRP area in 119 hours, or 5 days (see also Riehl et al. 1951). It should be emphasized that this is a rough estimate, since the average airflow may not be representative of airflow in specific cases, but consistent with the approximate nature of the subsequent analysis. [The above trajectory is not very different from that estimated by Neiburger (1960) based on data collected during the 1940s and 1950s. Neiburger’s trajectory that passes over Hawaii goes through 135° west (rather than 130° west) at the latitude of the FIRE research area. We believe this difference is not significant.] For simplicity, in the discussion that follows we use the expression “travel from FIRE to HaRP” to refer to travel along the above trajectory starting 6° west of the FIRE research area.

3. Measurements

The aircraft data were collected from the National Center for Atmospheric Research (NCAR) Electra during ascents and descents through and above the boundary layer. Our main focus is on three variables: liquid water potential temperature $\theta_l = \theta - (L/c_p)q_L$ (where $q_L$ is the liquid water mixing ratio, $L$ is the latent heat of water vapor, and $c_p$ is the specific heat of air at constant pressure), total water mixing ratio (vapor + liquid, $Q$, in g per kg of air), and ozone mixing ratio ($O_3$, in parts per billion by volume, ppb). The above variables are conserved during mixing, but can be affected by other internal processes. The $\theta_l$ can be changed by radiation and precipitation, $Q$ by precipitation, and $O_3$ by photochemical reactions.

The technique for measuring ozone during FIRE has been described by Pearson (1990). The measurements use chemiluminescence from the reaction of ozone with nitric oxide as an indicator of ozone concentration. An NCAR designed instrument similar to that of Pearson's, but with a reator described by Ridley (1992), was flown in HaRP. It was calibrated against a stable, slower response instrument that uses ultraviolet absorption for ozone detection. These instruments are described in detail by Luke et al. (1992) and Schillawski et al. (1991).

Other aircraft measurements include pressure, air temperature, liquid water from the FSSP, and moisture measured by a fast response Lyman-alpha hygrometer, which is calibrated against a dewpoint hygrometer for
Fig. 1. Surface wind stress vectors and divergence for August averaged over the years 1966–1985, and for August 1985 (Strickertz et al. 1992). The wind velocity has the same direction as the stress vector and a magnitude of $|\tau/(\rho C)|^{1/2}$, where $\tau$ is the magnitude of the wind stress (N m$^{-2}$), $\rho$ is the air density (1.2 kg m$^{-3}$), and $C$ is the drag coefficient ($1.5 \times 10^{-3}$). The circles marked $F$ and $H$ indicate the FIRE and HaRP research areas, which are about 3000 km apart.
each flight. The accuracy of the liquid water data from the FSSP is questionable but, for the soundings considered, the liquid phase contributes little (<2%) to the average total water mixing ratio in the boundary layer.

In addition to aircraft data there are satellite photographs. Examples from FIRE (Wylie 1993, private communication; see also Kloesel et al. 1988) are shown in Fig. 2. Of particular interest in this study is the breakup of solid cloud cover. The satellite photographs show that typically the cloud cover over the FIRE area becomes patchy and dissipates from one-quarter to one-half of the distance between FIRE and HaRP. In the HaRP area a solid cloud cover was not observed, but scattered clouds were common.

4. Comparisons of soundings from FIRE and HaRP

Figure 3 shows examples of soundings from HaRP and FIRE. The first two are clear-air soundings from HaRP; the last two are soundings from FIRE where the boundary layer was capped with a cloud deck 300 m and 150 m thick (soundings c and d, respectively). The capping inversion in the HaRP area, called the trade inversion, is considerably higher than the inversion in the FIRE area, consistent with many previous observations (e.g., Riehl et al. 1951). Soundings with very dry air and high ozone content above the capping inversion are shown in Figs. 3a and 3c, whereas Figs. 3b and 3d show soundings with large amounts of moisture and low ozone content above the inversion. Most

Fig. 2. Satellite photographs of cloud cover during FIRE on 11 July (top) and 16 July (bottom) 1987; visible: left, infrared: right (courtesy of Dr. Donald Wylie). The FIRE research area was at about 34°N, 124°W and the HaRP research area was at 20°N, 155°W (just outside the left edge of the figures).
Fig. 3. Examples of soundings from HaRP and FIRE showing liquid water potential temperature $\theta_l$ (K), total water mixing ratio (vapor + liquid, $Q$, in g per kg of air), and ozone mixing ratio ($O_3$, in ppb, or parts per billion by volume). The first two are clear-air soundings from HaRP; the last two are soundings from FIRE where the boundary layer was capped by a cloud deck 300 m and 150 m deep (c and d, respectively).
soundings in FIRE and HaRP fall between these extreme cases, more frequently tending toward the dry soundings in Figs. 3a and 3c.

The dashed line in Fig. 3b shows the moist adiabat for a lifted condensation level corresponding to the observed surface values: cloud base at 700 m with saturation water vapor mixing ratio of 14 g kg⁻¹. As can be seen, even though the capping inversion in HaRP is less pronounced than the inversion in FIRE, it is still sufficiently strong to prevent convection from rising through it. (The situation may change due to coastal effects, but here we consider only soundings away from the coast and on the upwind side of the Hawaiian island.) The air with high moisture content above the capping inversion in Figs. 3b and 3d must have been advected from a different area where at some earlier time the temperature stratification was favorable for deep convection.

In the following data summaries we include eight soundings from FIRE, which represent all flight days except the last two, where we observed signs of contamination by continental air from the coast (Paluch et al. 1992). For HaRP we use ten soundings upwind of Hawaii for which ozone data are available. In addition, three HaRP soundings for which ozone data were judged to be unreliable are included without the ozone data. In the FIRE soundings, the boundary layer was capped by a cloud layer 50–700 m deep. No cloud layer was present in the HaRP soundings. Except for a few small cloud patches, which probably were remnants of small cumuli, the HaRP soundings in this dataset are essentially clear-air soundings.

The characteristics of the lower free troposphere during FIRE and HaRP are summarized in Fig. 4. This figure shows θ̂, (or θ̂, since no clouds were present), Q, and O₃ averaged over 1-km altitude intervals for soundings from FIRE (short solid vertical lines) and HaRP (long dashed vertical lines), 2–5 km above sea surface. (We have used lines rather than dots to improve legibility; the lengths of the lines have no physical meaning.) At the level of 4–5 km the FIRE data are sparse because most soundings terminated at lower altitudes. The θ̂ profiles are very similar for the two field experiments, with a θ̂ increase with altitude of about 5 K km⁻¹. We expect that vertical mixing is negligible in this stably stratified region. The similarity of the θ̂ profiles from the two experiments indicates that radiative cooling of the lower free troposphere is balanced by a rise in the potential temperature due to subsidence along the FIRE–HaRP trajectory. In clear air, 2–5 km above sea surface, the radiative cooling rate is about 1.5 K/day (Roach and Slingo 1979; Fig. 4). For the observed θ̂ gradient of 5 K/km this cooling rate is balanced by a subsidence of 12 m h⁻¹, which corresponds to a divergence of 1.1 × 10⁻⁶ s⁻¹ at 3 km. This value is about the same as the mean surface divergence along the FIRE–HaRP trajectory in Fig. 1.

![Fig. 4. Liquid water potential temperature θ̂, total water mixing ratio (Q), and ozone mixing ratio (O₃) in the lower free troposphere versus height (Z). Data from FIRE (short solid vertical lines) and HaRP (long dashed vertical lines) represent individual aircraft soundings averaged over 1-km altitude intervals. (The length of the lines has no physical meaning.) Data from below the capping inversion are not included.](image-url)
For estimates in the following sections where a divergence is assumed, we use a value of $10^{-6} \text{s}^{-1}$ (constant with height), because this value fits well within the observed surface divergence range in the FIRE–HaRP area (Fig. 1), and because it balances the estimated radiative cooling of the lower free troposphere. Furthermore, a divergence of $10^{-6} \text{s}^{-1}$ or less keeps our estimates of mean surface moisture fluxes within a reasonable range. A divergence substantially larger than $10^{-6} \text{s}^{-1}$ requires an average surface moisture flux that is larger than the maximum observed during FIRE and HaRP, as will be discussed in the next section. It should be noted that the surface divergence increases towards the California coast (Fig. 1; Neiburger 1960). Air trajectories closer to the coast [as for example in Bretherton (1990)] encounter a stronger divergence than those along the FIRE–HaRP trajectory.

In both FIRE and HaRP, the water vapor and ozone mixing ratio in the lower free troposphere show much greater scatter than $\theta_L$ (Fig. 4). This is because $\theta_L$ is constrained by buoyancy forces, whereas water vapor and ozone are not. Figure 5 shows that water vapor and ozone are inversely correlated, indicating that in both field experiments ozone rich air had descended from higher altitudes where the air is very dry. [For a more detailed analysis of FIRE soundings see Paluch et al. (1992).] It is worth noting that in Fig. 5, there are no systematic differences between the data points from FIRE and HaRP that could be indicative of significant long-term changes or differences in calibrations of the ozone or humidity sensors between the two field experiments.

The characteristics of the troposphere below the capping inversion during FIRE and HaRP are summarized in Figs. 6, 7, and 8. The horizontal axis gives $Q$, $O_3$, and $\theta_L$; the vertical axis indicates the height of the capping inversion. To facilitate comparison, the values corresponding to the soundings in Figs. 3a–d are marked by letters a–d, respectively.

Figure 6 shows data collected in the well-mixed boundary layer within about 200 m of the sea surface, except for the two soundings in HaRP with the lowest $Q$ (and highest $O_3$) values, which were sampled 400 m above the surface. These could underestimate the surface $Q$ values (and overestimate $O_3$ values). As can be seen, during transit from FIRE to HaRP there is an increase in the surface values of $Q$ and $\theta_L$, and some decrease in surface $O_3$. Compared to the net changes there is much scatter in $O_3$ and $Q$, but relatively little in $\theta_L$. This is because for $\theta_L$ the changes are larger since surface fluxes and entrainment both act to increase $\theta_L$, whereas for $Q$ and $O_3$ the two processes counteract each other. Furthermore, above the capping inversion $Q$ and $O_3$ show relatively more variability than $\theta_L$ (Fig. 4); hence, entrainment will cause more variability in $Q$ and $O_3$ than in $\theta_L$.

Figure 7 shows the average values of $Q$, $O_3$, and $\theta_L$ below the capping inversion. In FIRE the air below the inversion was nearly well mixed (Paluch and Lenschow 1991), and thus, the average values do not deviate much from the surface values. In HaRP the well-mixed boundary layer was shallow and there were significant gradients in $Q$ and $\theta_L$ above it (as, for example, in Figs. 3a and 3b); hence, the average values below the capping inversion deviate from the surface values. During transit from FIRE to HaRP there is a net gain in $Q$ below the capping inversion, indicating that on the average moisture gain from sea surface is stronger than the moisture loss due to cloud-top entrainment and precipitation fallout (if it does occur). An overall decrease in $O_3$ is apparent, though the scatter is large, and there is some overlap between the FIRE and HaRP data. In contrast, the increase in $\theta_L$ is quite large, about 10 K or more, indicating that heat gained from sea surface, cloud top entrainment, and precipitation fallout dominate over radiative cooling.

Figure 8 shows values for $Q$, $O_3$, and $\theta_L$ integrated with height up to 3 km. These values represent the area to the left of the sounding curves. We have chosen 3 km as the upper integration level because in all cases this level is well above the capping inversion. There the air is very stably stratified, and turbulent fluxes through the top of the layer can be ignored. The differences between the integrated areas in FIRE and HaRP soundings are not affected by mixing across the inversion, but represent the net gain or loss due to sea surface fluxes, precipitation fallout, and...
subidence, as well as radiative cooling and warming for $\theta_L$. The gain in the integrated moisture in the HaRP data indicates that moisture gain from the sea surface is larger than loss of moisture due to subidence and precipitation fallout. The decrease in the integrated ozone indicates that ozone loss due to photochemical reactions and deposition at sea surface dominates ozone gain due to subidence. There is no significant change in the integrated $\theta_L$, implying that the gain in $\theta_L$ from sea surface, subidence, and precipitation fallout is counterbalanced by radiative cooling.

The capping inversion is considerably higher in HaRP than in FIRE, and its height is quite variable. This can be expected to reflect the net balance between the rates of entrainment and subidence and the variability in wind speed, or transit time from FIRE to HaRP.

5. Mean fluxes and cloud growth

We estimate the mean fluxes by first comparing the averages from the FIRE and HaRP datasets, then comparing individual soundings that have similar upper-level structure: the dry upper-level soundings shown in Figs. 3a and 3c, and the moist upper-level soundings shown in Figs. 3b and 3d.

As shown in the Appendix, the mean fluxes can be estimated from the integrated areas in Fig. 8:

$$\langle w'q' \rangle_0 - \langle w'q' \rangle_Z \approx \frac{\overline{Q_{Hz}Z} - \overline{Q_{rZ}}}{T}$$

$$- \left[ \overline{Q_{Hz}Z} - \overline{Q_{Hz}Z} + Q_{zrZ} - \overline{Q_{rZ}} \right] D/2,$$

where subscripts $H$ and $F$ refer to HaRP and FIRE soundings, $0$ and $Z$ to the surface level and the upper integration level, $T$ is the transit time from FIRE to HaRP, $w'q'$ the total water flux including precipitation, $D$ is divergence, the rate of subidence $= Dz$ (where $z$ is the altitude), and

$$\overline{Q_{rZ}} = \int_0^Z Q_r dz.$$ 

For data in Fig. 8 we have chosen the integration limit $Z$ at 3 km, which is well above the capping inversion where the vertical fluxes are negligible, so that $w'q_Z$ $\approx 0$.

The above equation is based on the assumptions that differential horizontal advection is negligible and that the properties of the upper-level air do not change drastically, so that we can estimate the mean contribution
from subsidence by taking the average of the integrated soundings at the beginning and end of the trajectory.

a. Average soundings

1) MOISTURE FLUX

The difference between the HaRP and FIRE soundings in the integrated moisture content over the lower 3 km (average values from Fig. 8) is (26.2 - 13.7) g kg\(^{-1}\) km = 12.5 g kg\(^{-1}\) km. This value is the net result of moisture gain from the sea surface and moisture loss due to precipitation and subsidence. If there is no moisture flux across the 3-km level and no subsidence, then during the 5-day (119 h) transit, this corresponds to an average net surface moisture flux (including precipitation) of 12 500 g kg\(^{-1}\) m/(3600 x 119 s) = 0.03 g kg\(^{-1}\) m s\(^{-1}\). In the presence of subsidence due to 10 x 10\(^{-6}\) s\(^{-1}\) divergence, the resulting net surface moisture flux is 0.04 g kg\(^{-1}\) m s\(^{-1}\) (with mean \(Q_{zf}\) \sim Q_{zh} \sim 2 g kg\(^{-1}\) at 3 km, from Fig. 4b). The estimated moisture fluxes are somewhat high, but not unrealistic. Sea surface moisture fluxes up to 0.03 and 0.04 g kg\(^{-1}\) m s\(^{-1}\) were observed in FIRE and HaRP, respectively. [All fluxes quoted here and elsewhere in this paper were obtained using a 5-km wavelength high-pass filter, which can be expected to underestimate the total flux by about 10% (Lenschow et al. 1993).] If we assume a divergence of 3.0 x 10\(^{-6}\) s\(^{-1}\), then this gives a mean surface flux of 0.07 g kg\(^{-1}\) m s\(^{-1}\) (or more, if the precipitation flux is significant). This value is much higher than the maximum observed during FIRE and HaRP.

2) OZONE FLUX

The changes in ozone (\(O_3\)) integrated over the first 3 km are (117 - 73) ppb km = 44 ppb km, which in the absence of subsidence gives an average downward ozone flux at the surface of 0.10 ppb m s\(^{-1}\) during the 5-day transit. If we assume no photochemical ozone destruction and a typical surface ozone concentration of 20 ppb, the above flux results in an average ozone surface deposition velocity of 0.5 cm s\(^{-1}\) (and more if there is subsidence). This estimate is an order of magnitude larger than the value for deposition velocity measured by Kawa and Pearson (1989) using eddy correlation flux measurements from DYCOS.

Evaluating the mean ozone budget, Kawa and Pearson also estimated that the boundary layer had a net ozone sink of 0.07 ± 0.11 ng m\(^{-3}\) s\(^{-1}\), or 3.0 ± 4.7 ppb day\(^{-1}\). The ozone sink is thought to be the result of ozone destruction by ultraviolet radiation in a low NO\(_x\) environment (Liu et al. 1983; Kawa and Pearson 1989; Johnson et al. 1990). We shall now estimate the strength of this ozone sink from the present data, assuming that the average ozone deposition velocity is 0.026 cm s\(^{-1}\) as reported by Kawa and Pearson. For a surface ozone concentration of 20 ppb, the flux due to surface deposition is 0.005 ppb m s\(^{-1}\), only 5% of the
estimated average flux. Hence most of the ozone loss must be attributed to the ozone photochemical sink. Assuming that this sink is below the capping inversion and using 1300 m for the mean inversion height gives an ozone sink of 6.5 ppb day\(^{-1}\) in the absence of subsidence, and 8.2 ppb day\(^{-1}\) in the presence of subsidence due to 10\(^{-5}\) s\(^{-1}\) divergence (with \(O_3 = 10^{-5}\) s\(^{-1}\) and a mean inversion height of 1.3 km, then the average entrainment velocity is 14 m h\(^{-1}\) or 0.39 cm s\(^{-1}\).

To infer the effects of cloud-top entrainment we need to know the mean jumps of \(\theta_L\), moisture and ozone across the capping inversion. However, the ozone and humidity structure above the inversion are too variable to deduce mean jumps that could be viewed as representative of the HaRP and FIRE datasets. For this reason we do not attempt to estimate the effects of entrainment using mean values. Instead, we shall consider some specific cases. Because of the strong capping inversion a typical dry upper-level sounding cannot be transformed into a wet upper-level sounding and vice versa, except through advection. To minimize possible advective effects we shall estimate fluxes by comparing soundings from FIRE and HaRP that have a similar upper-level structure: first the dry upper-level soundings shown in Figs. 3a and 3c, then the moist upper-level soundings shown in Figs. 3b and 3d.

b. Dry upper-level soundings

1) Moisture flux

The difference between the HaRP and FIRE soundings in the integrated moisture content (a and c in Fig.
8) is \((24.5 - 9.5) \text{ g kg}^{-1} \text{ km} = 15 \text{ g kg}^{-1} \text{ km}\). This value is the net result of moisture gain from the sea surface, and moisture loss due to precipitation and subsidence. If there is no moisture flux across the 3-km level and no subsidence, then during the 5-day transit this corresponds to an average surface moisture flux (including precipitation) of \(0.035 \text{ g kg}^{-1} \text{ m s}^{-1}\). In the presence of subsidence due to \(10^{-6} \text{ s}^{-1}\) divergence and with \(Q_{zf} \approx Q_{zh} \approx 0\) at 3 km (Figs. 3a and 3c), the resulting surface moisture flux (including precipitation) is \(0.052 \text{ g kg}^{-1} \text{ m s}^{-1}\).

2) Ozone flux

The change in ozone \((O_3)\) integrated over the first 3 km is \((121.8 - 92.5) \text{ ppb km} = 29.3 \text{ ppb km}\), which in the absence of subsidence gives an average downward ozone flux at the surface of 0.068 ppb m s\(^{-1}\) during the 5-day transit. If we assume no photochemical ozone destruction and a typical surface ozone concentration of 20 ppb, the above flux results in an average ozone surface deposition velocity of 0.34 cm s\(^{-1}\), which again is much larger than the value for deposition velocity estimated by Kawa and Pearson (1989).

Using 1500 m for the mean height of the capping inversion, and assuming no subsidence and an average ozone deposition velocity of 0.026 cm s\(^{-1}\), gives an ozone sink of 3.6 ppb day\(^{-1}\). With subsidence due to \(10^{-6} \text{ s}^{-1}\) divergence the ozone sink is 6.9 ppb day\(^{-1}\) (where \(O_{3f} \approx O_{3h} \approx 55 \text{ ppb}\) at 3 km, from Figs. 3a and 3c).

It could be argued that large uncertainties are introduced when comparisons are made using ozone data from different years measured by different instruments (though both use the same technique). In response to this we can use another approach: instead of using the FIRE sounding (Fig. 3c), we can extrapolate the HaRP sounding (Fig. 3a) backwards in time, making only the assumptions that 5 days earlier the capping inversion was at about 1 km and that the mean ozone concentration above the inversion was unchanged. This extrapolation results in an ozone profile that is very similar to the FIRE sounding in Fig. 3c; hence, the ozone flux estimated this way would not be very different from the previous estimate.

3) Liquid water potential temperature flux

The values for \(\theta_e\) integrated over the first 3 km are essentially the same in both soundings. (As can be seen in Fig. 8, this is not unusual.) The increase in \(\theta_e\) due to sea surface temperature flux, subsidence, and precipitation fallout must have balanced the decrease in \(\theta_e\) due to radiative cooling.

4) Entrainment across the capping inversion

During transit from FIRE to HaRP, the height of the capping inversion has increased by about 1.3 km, which corresponds to an average rise rate of 11 m h\(^{-1}\) or 0.30 cm s\(^{-1}\). If we assume subsidence due to a mean divergence of \(10^{-6} \text{ s}^{-1}\), then the average cloud-top entrainment velocity is 0.45 cm s\(^{-1}\).

5) Will the cloud layer grow or dissipate?

Using the estimated mean time rates of change for \(Q\) and \(\theta_e\), we shall now examine the sensitivity of the predicted cloud-layer growth or dissipation rates to the assumed turbulence structure below the capping inversion, first assuming a well-mixed boundary layer extending up to the capping inversion (as in a mixed-layer model), then a cloud layer decoupled from the surface layer.

We estimate the tendency for the cloud layer to grow or dissipate from the mean cloud-base and cloud-top rates of rise. Changes in cloud-base pressure \((P)\) can be calculated from (Paluch and Lenschow 1991):

\[
\frac{d \ln P}{dt} = \frac{L \frac{dT}{dt} - d \ln q_s}{R_T \frac{dT}{dt}} \left[ 1 + \frac{LR}{gR_T} \frac{\partial T}{\partial z} \right]^{-1},
\]

where \(T\) is the temperature, \(q\) is the water vapor mixing ratio, subscript \(s\) denotes water saturation at cloud base, \(L\) is the latent heat of vaporization, \(g\) is gravitational acceleration, and \(R\) and \(R_T\) are the gas constants for air and water vapor, respectively.

Using the hypsometric equation for conditions considered here (cloud base at about 11°C and 950 mb) this reduces to

\[
\frac{dz}{dt} \approx 124 \frac{d\theta_e}{dt} - 216 \frac{dQ}{dt}
\]

where the cloud-base altitude \(z\) is in meters and \(Q\) is in grams per kilogram. (Here \(Q\) and \(\theta_e\) are evaluated at cloud base. In a well-mixed layer they correspond to the average values in the layer.)

(i) Mixed layer extending up to the capping inversion

In the case of the well-mixed boundary layer, we can obtain the net time rates of change directly from the mean values below the capping inversion and the inversion height. From Fig. 7, the average \(dQ/dt = (10.7 - 8.9)/119 \text{ g kg}^{-1} \text{ h}^{-1} = 0.015 \text{ g kg}^{-1} \text{ h}^{-1}\), and the average \(d\theta_e/dt = (299.0 - 288.3)/119 \text{ K h}^{-1} = 0.99 \text{ K h}^{-1}\). Substituting these values in (2) gives a cloud-base rise rate of 8 m h\(^{-1}\). As the cloud top rises along with the capping inversion at an average rate of 11 m h\(^{-1}\), the cloud layer will increase in thickness at an average rate of 3 m h\(^{-1}\). (We note that this estimate is independent of the assumed rates of subsidence and entrainment across the inversion.)

(ii) Decoupled cloud layer

While surface heating and cloud-top radiative cooling promote turbulence in the boundary layer, short-
wave heating in cloud during the day and cooling due to evaporation of precipitation below cloud base tend to suppress turbulence by stabilizing the temperature profile. When the latter processes dominate, the cloud layer can become decoupled from the well-mixed layer near surface. Decoupled cloud layers were frequently observed during ASTEX, and they occurred at night as well as during the day. [For more discussion of observations and/or modeling of decoupling, see, for example, Turton and Nicholls (1987); Duyunkerke (1989); Betts (1990); Higlett (1991); Paluch and Lenschow (1991); Wang and Wang (1994).] Without speculating how the decoupling has evolved, we consider here a rather common decoupling situation: a well-mixed cloud layer extending up to the capping inversion that is decoupled from the surface layer by a stable layer below cloud base. The cloud layer is not directly affected by surface fluxes, but the effects of in-cloud radiation exchange and cloud-top entrainment here are more pronounced since they are distributed over the cloud depth only.

Earlier we estimated a 0.45 cm s\(^{-1}\) entrainment velocity across the capping inversion assuming subsidence due to 10\(^{-6}\) s\(^{-1}\) divergence. In Figs. 3a and 3c the average moisture jump across the inversion is \(\approx 8\) g kg\(^{-1}\), which corresponds to an entrainment moisture flux of 0.036 g kg\(^{-1}\) m s\(^{-1}\). For the FIRE sounding in Fig. 3c, the cloud depth is 300 m. The entrainment moisture flux \(\frac{dQ}{dt} = -0.43\) g kg\(^{-1}\) h\(^{-1}\) averaged over the cloud depth. Assuming an average \(\theta_e\) jump across the capping inversion of 10 K (Figs. 3a and 3c), the entrainment \(\theta_e\) flux is \(-0.045\) K m s\(^{-1}\), so that the average \(\frac{d\theta_e}{dt}\) due to entrainment in the cloud layer is 0.54 K h\(^{-1}\). For an adiabatic cloud layer 300 m thick, the longwave radiative cooling rate is about 0.5 K h\(^{-1}\), and shortwave solar heating rate up to 0.9 K h\(^{-1}\) at noon, both averaged over the cloud depth. [These values were calculated from the model described by Siems et al. (1993), with input parameters adjusted to fit the soundings.] This gives a net \(\frac{d\theta_e}{dt} = 0.04\) K h\(^{-1}\) at night, and a net \(\frac{d\theta_e}{dt} = 0.94\) K h\(^{-1}\) at noon. Substituting these values in (2), we find that the cloud base will rise rapidly—at a rate of 100 m h\(^{-1}\) during the night, and 210 m h\(^{-1}\) at noon, which should lead to the dissipation of the cloud layer. (As the cloud layer thins, the cloud-base rate of rise will tend to increase, at least during the night, mainly because the effects of cloud-top entrainment are distributed over a shallower layer.) Had we assumed a stronger subsidence, then for the same inversion rate of rise, the entrainment rate across the inversion would have to be higher, leading to higher estimates of cloud-base rates of rise.

The differences in the cloud-base rise rates during night and day are less pronounced if we assume a diurnal variation in the cloud-top entrainment rate. If the entrainment rate is 40% above average during the night and 40% below average during the day (as in Bougeault 1985, Fig. 19), then the cloud-base rise rate is 160 m h\(^{-1}\) at night and 150 m h\(^{-1}\) at noon.

While not much confidence can be placed on the specific numerical values of cloud-base rates of rise, all the above assumptions lead to dissipation of the stratus layer within a few hours or less. Once the cloud layer has dissipated, turbulence will decrease, since there is no cloud-top radiative cooling to supply energy for mixing. However, this does not imply that clear sky will follow. After sufficient moisture from the sea surface has accumulated below the stable layer, moist convection (typically in the form of scattered cumuli or cloud streets) may penetrate and locally destroy the stable layer (and the decoupling), but this is outside the scope of the present discussion.

c. Moist upper-level soundings

We now consider the two soundings in Figs. 3b and 3d. The large variations in ozone and moisture above the capping inversion can be expected to introduce large uncertainties in the estimates, but as will be seen, the results are still of the same order of magnitude as in the previous case.

1) Moisture Flux

The gain in \(Q\) integrated over the first 3 km (b and d in Fig. 8) is (25.6 - 19.7) g kg\(^{-1}\) km = 5.9 g kg\(^{-1}\) km. This corresponds to an average net moisture flux of 0.014 g kg\(^{-1}\) m s\(^{-1}\). Again this value represents the combined effects of sea surface flux, precipitation fallout, and subsidence. In the presence of subsidence due to 10\(^{-6}\) s\(^{-1}\) divergence the resulting surface moisture flux (including precipitation) is 0.03 g kg\(^{-1}\) m s\(^{-1}\) (with \(Q_{ZH} \approx 5\) g kg\(^{-1}\) and \(Q_{ZF} \approx 4\) g kg\(^{-1}\) at 3 km, from Figs. 3b and 3d).

2) Ozone Flux

The changes in ozone (\(O_3\)) integrated over the first 3 km are (76 - 60) ppb km = 16 ppb km, which in the absence of subsidence gives an average downward ozone flux at the surface of 0.038 ppb m s\(^{-1}\). Assuming a typical surface ozone concentration of 20 ppb, the above flux results in an average ozone surface deposition velocity of 0.19 cm s\(^{-1}\). If the average ozone deposition velocity is 0.026 cm s\(^{-1}\), as reported by Kaw and Pearson (1989), and the mean inversion height is 1 km, then in the absence of subsidence there should be an ozone sink of about 2.8 ppb day\(^{-1}\), and not much higher in presence of subsidence, since the mean ozone concentration above the inversion in Figs. 3b and 3d is not very different from that below the inversion.

3) Liquid Water Potential Temperature (\(\theta_L\)) Flux

The gain in \(\theta_L\) integrated over the first 3 km is (912.7 - 908.4) K km = 4.3 K km, which corresponds to an
average net flux of 0.01 K m s\(^{-1}\), representing the net effects of sea surface temperature flux, subsidence, precipitation fallout, and radiation.

4) **Entrainment Across the Capping Inversion**

Here the capping inversion has risen by about 0.8 km, which corresponds to an average rise rate of 6.7 m h\(^{-1}\) or 0.2 cm s\(^{-1}\). In the presence of subsidence due to 10\(^{-6}\) s\(^{-1}\) divergence, the average cloud-top entrainment velocity is 0.3 cm s\(^{-1}\).

5) **Will the Cloud Layer Grow or Dissipate?**

(i) **Mixed layer extending up to the capping inversion**

From Fig. 7, the average \(dQ/dt = (12 - 9)/119\) g kg\(^{-1}\) h\(^{-1}\) = 0.026 g kg\(^{-1}\) h\(^{-1}\), and the average \(d\theta_L/dt = (298.2 - 287.3)/119\) K h\(^{-1}\) = 0.09 K h\(^{-1}\). Substituting these values in (2) gives a cloud-base rise rate of 6 m h\(^{-1}\). Since the capping inversion rises at an average rate of 7 m h\(^{-1}\), the depth of the cloud layer should increase at an average rate of 1 m h\(^{-1}\). (As in the previous dry upper-level soundings, this estimate is independent of the assumed rates of subsidence and entrainment across the inversion.)

(ii) **Decoupled cloud layer**

Earlier we estimated a 0.3 cm s\(^{-1}\) entrainment velocity in the presence of subsidence due to 10\(^{-6}\) s\(^{-1}\) divergence. The fluctuations in \(Q\) above the capping inversion in Figs. 3b and 3d suggest that the moisture jump across the inversion will vary as the inversion rises. In Fig. 3b there is a dry layer just above the inversion. In Fig. 3d there is also a dry layer near the inversion, but this layer is actually about 20 mb above the inversion. Thus, the moisture jump across the inversion can be expected to vary between 0 and -6 g kg\(^{-1}\). Let us assume a rather modest value of -2 g kg\(^{-1}\) for the average moisture jump across the inversion. This corresponds to an entrainment moisture flux of 0.006 g kg\(^{-1}\) m s\(^{-1}\). The cloud depth in the sounding in Fig. 3d is 150 m, which gives \(dQ/dt = -0.14\) g kg\(^{-1}\) h\(^{-1}\) averaged over the cloud depth. Assuming a 10 K \(\theta_L\) jump across the inversion, the entrainment \(\theta_L\) flux is -0.03 K m s\(^{-1}\), and the average \(d\theta_L/dt\) due to entrainment is 0.72 K h\(^{-1}\). For an adiabatic cloud layer 150 m thick, the longwave radiative cooling rate is about -0.9 K h\(^{-1}\), and shortwave solar heating rate up to 0.8 K h\(^{-1}\) at noon, both averaged over the cloud depth. This gives a net \(d\theta_L/dt = -0.18\) K h\(^{-1}\) at night, and a net \(d\theta_L/dt = 0.62\) K h\(^{-1}\) at noon. Substituting these values in (2), we find that the cloud base will rise at 9 m h\(^{-1}\) during night, and at 110 m h\(^{-1}\) at noon.

If we assume that the cloud-top entrainment rate is 40% above average during the night and 40% below average during the day, as in the previous case, then the cloud base rise rate is 57 m h\(^{-1}\) at night and 60 m h\(^{-1}\) at noon. These rise rates are almost three times slower than in the dry upper-level sounding case, but here too they can be expected to cause the 150-m cloud layer to dissipate.

6. **Discussion**

Aside from the estimated fluxes and other quantities that could be useful for making comparisons with numerical model results, there are two main points that have emerged from the FIRE–HaRP sounding comparisons.

1) The estimates indicate that stratus growth and dissipation are very sensitive to the assumed turbulence structure below the capping inversion. For the average conditions in the two types of soundings considered here, a mixed-layer (or bulk) model that assumes a well-mixed boundary layer up to the capping inversion predicts a solid cloud layer extending all the way to Hawaii, whereas a model that allows decoupling predicts dissipation of the cloud layer. This suggests that cloud dissipation here is due to a reduction of turbulent mixing below the capping inversion. Further insight may be gained by considering the opposite situation: let us assume that in the HaRP region there is sufficient turbulent energy to thoroughly mix the layer below the capping inversion, so that this layer has a constant \(Q\) and \(\theta_L\) equal to the average values (Fig. 7). In this situation, all the HaRP soundings have enough moisture to form a stratus layer, ranging in depth from 150 to 1000 m, shown in Fig. 9 (but, as discussed earlier, there were no stratus layers in our HaRP soundings).

In recent years much effort has been directed toward modeling the growth and dissipation of marine stratus within the framework of the mixed-layer model. In this type of model the dissipation of stratus is caused by drying of the mixed layer due to entrainment of dry air.
from above the inversion, and this has led to research focused on cloud-top entrainment instability as the controlling factor for stratus breakup. Present observations suggest that in the FIRE–HaRP area, stratus dissipation is not caused by drying of the mixed layer, but rather by a reduction of turbulent mixing within that layer. Hence, the mixed-layer model cannot be expected to predict realistic dissipation of stratus clouds in this area. Betts (1989), using an idealized diagnostic model for mixed and partially mixed boundary layers, reached a similar conclusion. Thus, for predicting stratus dissipation, models are needed that have vertical resolution and that can deal with the production and decay of turbulence. This is a formidable problem for numerical modeling. Where or when a cloud layer will dissipate and what type of cloud formation may follow should be sensitive to the moisture profile that forms during the reduction of turbulent mixing (for example, whether moisture decreases with height gradually or in a stepwise fashion as is characteristic of two or more decoupled well-mixed layers). The reduction of turbulent mixing can be initiated by shortwave heating in clouds during the day or evaporation of precipitation below cloud base, both of which stabilize the temperature profile. The process of precipitation formation in shallow cloud layers is presently not well understood, and this poses yet another major problem for modeling stratus breakup.

2) The differences in ozone concentrations below the capping inversion in HaRP and FIRE suggest the presence of a 3–8 ppb day$^{-1}$ ozone sink, corresponding to an ozone lifetime (time for 1/e decrease) of 3–9 days. This sink is about an order of magnitude more effective in removing ozone than deposition at the sea surface. The present estimate of ozone photochemical lifetime in the moist layer is within the range of previous estimates for the southeastern Pacific in the summer by Kawa and Pearson (1989), who report a photochemical ozone sink of 3.0 ± 4.7 ppb day$^{-1}$, or a lifetime >4 days, and estimates for the equatorial Pacific by Liu et al. (1983), who infer an ozone lifetime of >1 day and <16 days from observations, and about 8 days from model calculations. More recent calculations using a model similar to that of Liu et al. (1983) and conditions characteristic of the FIRE–HaRP area give ozone lifetimes in the range of 7–13 days, depending on cloud cover and the moisture content. More detailed analysis of ozone loss during FIRE and ASTEX will be the subject of our next paper.

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APPENDIX

Estimates of Fluxes from Soundings

Some definitions:

\[ z \text{ altitude} \]
\[ Q_H \text{ total water mixing ratio in HaRP sounding} \]
\[ Q_R \text{ total water mixing ratio in FIRE sounding} \]
\[ T \text{ transit time from FIRE to HaRP} \]
\[ w'q' \text{ total water flux including precipitation} \]
\[ Z \text{ upper integration level} \]
\[ 0 \text{ surface level} \]
\[ w \text{ vertical velocity due to subsidence} \]
\[ D \text{ divergence (assumed constant with height), so that } w = -zD. \]

Consider an air mass traveling along the FIRE–HaRP trajectory. The difference between the vertically integrated initial and final Q profiles can be expressed as

\[
\int_0^Z Q_H dz - \int_0^Z Q_R dz = \int_0^T \left[ \int_0^z \frac{d}{dz} \left( \frac{dQ}{dz} \right) dz \right] dt. \quad (A1)
\]

Assuming no differential horizontal advection, the rate of change in Q can be written in terms of the net total water flux vertical divergence and subsidence. Thus, the right side of (A1) becomes

\[
\int_0^T \left[ \int_0^z \left( \frac{d}{dz} \left( w'q' - w \frac{dQ}{dz} \right) \right) dt \right] dz

= \int_0^T \left[ \int_0^z \left( - \frac{d}{dz} w'q' + zD \frac{dQ}{dz} \right) dz \right] dt

= \int_0^T \left[ w'q'_0 - w'q'_z + D \left( ZQ_z - \int_0^z Q dz \right) \right] dt

= \left( \langle w'q' \rangle_0 - \langle w'q' \rangle_z + \langle DZq \rangle \right)

- \left( D \int_0^z Q dz \right) T, \quad (A2)
\]

where the angular brackets denote a time average, subscripts 0 and Z refer to the altitude where the quantities are evaluated, and integration by parts has been used to evaluate the divergence term (assuming the vertical profiles can be smoothed into monotonic functions of altitude).

For similar upper-level soundings, the time averages in the subsidence terms can be estimated using the average of the two soundings. Thus, we can write

\[
\langle w'q' \rangle_0 - \langle w'q' \rangle_z = \frac{[Q_HZ - Q_RZ]}{T}

- \frac{[Q_HZ - Q_RZ + Q_RZ - Q_HZ]}{D}, \quad (A3)
\]

where \( Q_HZ = \int_0^Z Q_H dz \). If the integration limit Z is well above the capping inversion where the vertical turbulent fluxes are negligible, then \( \langle w'q' \rangle_z = 0 \).
Equations similar to (A3) can be used for relating $O_3$ and $\theta_r$ fluxes to the integrated areas, except $O_3$ includes a term for photochemical reactions and $\theta_r$ includes a radiation term, which cannot be readily evaluated, since it depends on cloud cover.

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