# CONVECTION IN STRATOCUMULUS-TOPPED ATMOSPHERIC BOUNDARY LAYERS

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## 1. Introduction

The dynamics of atmospheric convection are greatly enriched by the phase changes of water that can occur as air circulates. These phase changes are vital to the development of severe thunderstorms and many other forms of deep moist convection. They are equally important in the development of convecting cloud-topped boundary layers (CTBLs). CTBLs are less spectacular than deep moist convection, but they are ubiquitous and hence climatically important, covering the majority of Earth's oceans and much of its land surface.

CTBLs extend from the surface to a typical depth of 500-2000 m. The cloud may be a solid stratus layer a few hundred meters thick, or shallow cumuli up to 1500 m thick. The convection is maintained by an intimate set of feedbacks involving condensation into cloud, radiative cooling, surface fluxes, precipitation, and entrainment of air from above the boundary layer. These feedbacks have proven challenging to represent in numerical global circulation models for climate simulation, because the convective circulations are not resolved in such models, and their effects must be parameterized.

In this article, we examine the global distribution and climatic importance of boundary layer cloud, then discuss the fascinating dynamics that produce different types of CTBLs. Our main focus will be on the formation, maintenance and breakup of CTBL's capped by marine stratus or stratocumulus cloud layers, which typically cover one quarter of Earth's surface (Randall et al. 1985).



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## 2. Global distribution and importance of boundary layer cloud

Perhaps our most reliable climatology of boundary layer cloud comes from routine hourly weather reports from surface observers at weather observing sites and on commercial ships. They classify the cloud type and estimate the fraction of the sky (in eighths) covered by cloud. Although individual observations are imprecise, millions of routine surface cloud observations have now been archived. Satellite cloud climatologies are also useful, but usually cannot detect low-lying clouds underneath a higher cloud layer.

Figure 1 shows the annually averaged cloud amount (the product of cloud fraction and frequency of occurrence) for boundary layer stratus (or layer) clouds. These cloud layers are typically 100-500 m thick, with a cloud base anywhere from the surface to 1500 m, and tend to be nonprecipitating. Over much of the midlatitude oceans and parts of the eastern subtropical oceans, stratus cloud cover exceeds 50%. Klein and Hartmann (1993) showed that the cloud cover in these regions is highest when the sea-surface is coldest compared to the air above the boundary layer, which tends to occur in the summertime. In some parts of the Aleutian Islands, the average stratus cloud cover in June, July and August is 90%...a dreary sky indeed. Over land, there is much less stratus cloud due to the lesser availability of surface water. In most of the tropical and subtropical oceans, stratus clouds are rare. Instead, shallow 'trade' cumulus clouds are seen almost all the time. Even in these regions, trade cumulus cloud amounts do not exceed 10-20%, because they only cover a small part of the sky.

Both of these cloud types are important to global climate. Trade cumulus clouds help convect moist air upward from the ocean surface, greatly enhancing the amount of evaporation that occurs in the subtropics. The moistened air is drawn into a zone of persistent deep convection, the intertropical convergence zone, where much of the moisture is precipitated out. The resulting latent heating drives the entire tropical circulation. Tiedtke *et al.* (1988) found that a better parameterization of boundary layer cumulus convection in a weather forecast model considerably improved their representation of the strength of the mean tropical circulation.

Stratus clouds are also usually associated with convection. Often a distinct cellular structure associated with the convection can be discerned in the cloud brightness and cloud base height when the cloud layer is viewed from below. In this case, the clouds are usually called stratocumulus rather than stratus, but we will not concern ourselves with this technical distinction. The convection associated with stratus-capped boundary layers affects the surface temperature, relative humidity, and surface heat and moisture fluxes over much of Earth's surface.

Perhaps the most profound impact of stratus clouds on climate is due

Annual Stratus Cloud Amount

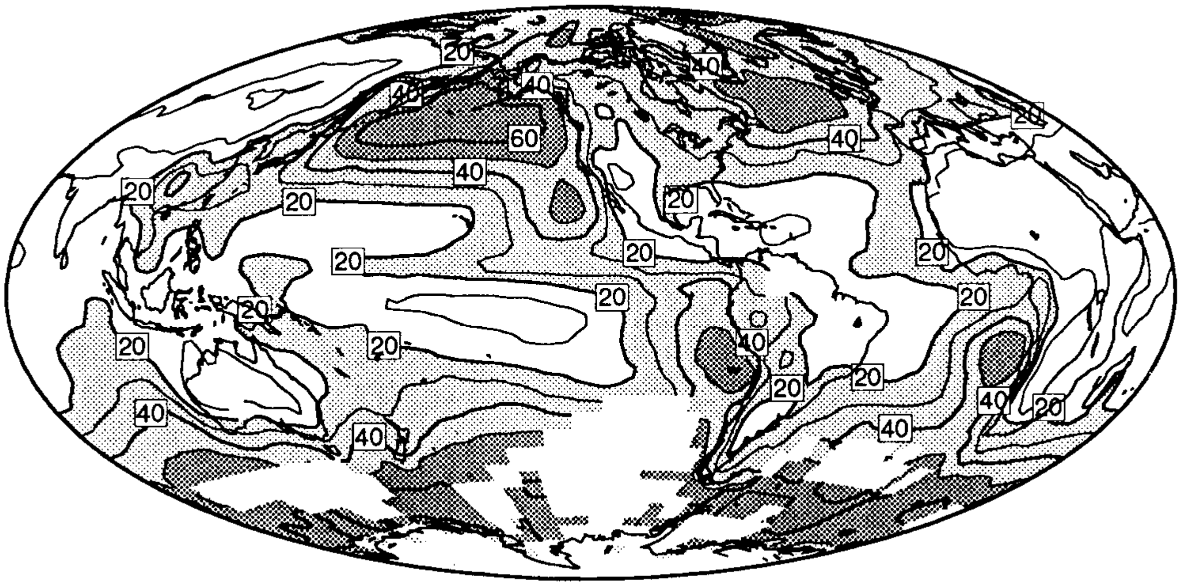



*Figure 1.* Annual average boundary layer stratus, stratocumulus and fog cloud amount in percent as seen by surface observers, from Klein and Hartmann (1993).

to their effect on Earth's radiation budget. Liquid water clouds as little as 100 m thick are almost opaque to infrared radiation. Therefore, they have a 'greenhouse' effect, absorbing upwelling infrared blackbody radiation emitted by the underlying surface, while radiating less infrared radiation upward because they are typically at a colder temperature than the surface. This effect is not that large for boundary layer clouds, which are very low in the atmosphere and hence nearly as warm as the surface, but is considerable for high cirrus clouds. Clouds also have an 'albedo' effect – they usually reflect solar radiation more efficiently than the underlying surface, so they increase the amount of solar radiation reflected back to space. Boundary layer stratus cloud layers have typical albedos of 30-70%.

Both of these effects have been measured from satellites, which can measure the 'cloud radiative forcing', which is defined as the difference between the combined upwelling infrared and visible radiance at times when a given location is cloud-free and its average at that location over all times. Figure 2 shows the annually averaged cloud radiative forcing. It bears a strong resemblance to the annually averaged stratus cloud amount. Because of their global coverage, boundary layer stratus clouds have a large albedo effect which dominates their greenhouse effect to create negative cloud radiative forcing, i. e. increased radiation of energy back to space. This helps cool the Earth. Hence, boundary layer stratus clouds can be thought of giant

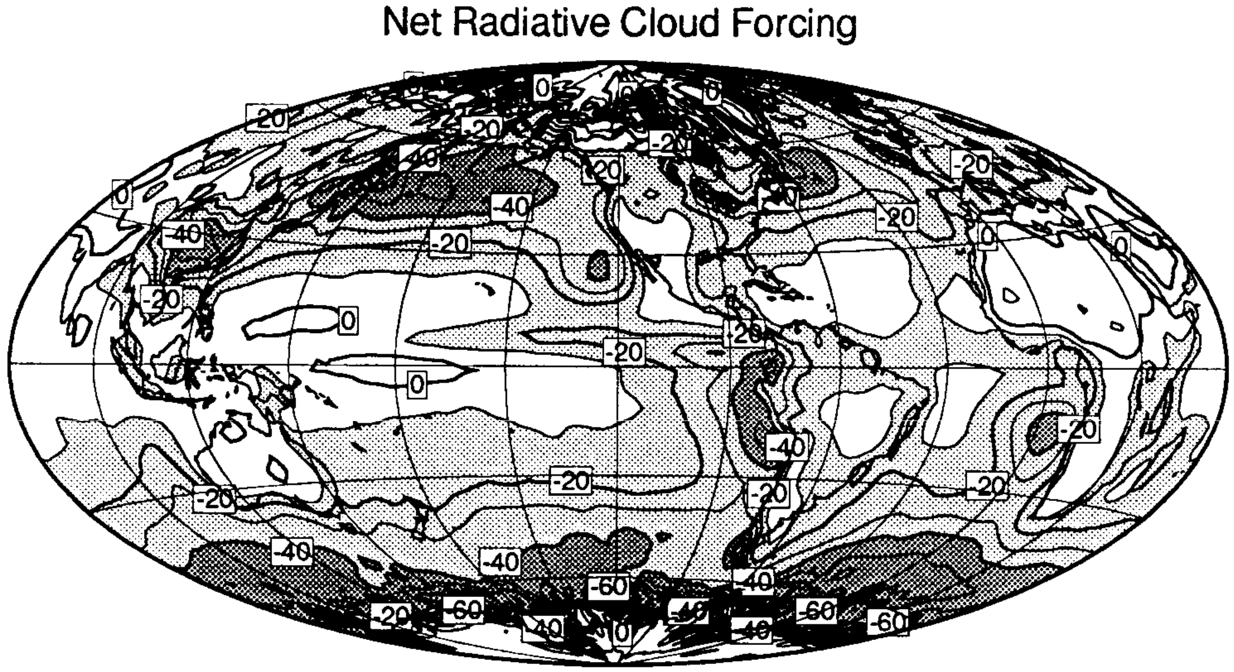



*Figure 2.* Satellite derived annual average cloud radiative forcing (W $m^{-2}$) for 1985-1986, from Klein and Hartmann (1993).

natural refrigerators for Earth and particularly its colder oceans. Cirrus clouds, by contrast, have considerable but almost cancelling albedo and greenhouse effects, and do not contribute significantly to the global pattern of net cloud radiative forcing. Over the midlatitude oceans, multilevel cloud systems associated with extratropical cyclones are common and may augment the negative stratus cloud radiative forcing, but even over these regions, roughly half of the cloud visible from space is in the boundary layer.

In the rest of this chapter, we will focus on marine CTBLs because they cover a much larger area than CTBLs over land and involve a more limited set of feedbacks. We have mentioned that stratus clouds are prevalent over the colder oceans but give way to shallow cumulus clouds over the warmer oceans. How do boundary layer dynamics help to produce this pattern? The answers will prove quite subtle, but we can learn more by looking at how the vertical structure of marine CTBLs varies between different locations.

Figure 3 shows composite soundings from four field experiments that studied marine subtropical and tropical CTBLs (Albrecht et al. 1995). The experiments were conducted over locations with very different sea-surface temperature (SST). The first variable plotted is the virtual potential temperature $\theta_v$. Vertical gradients in $\theta_v$ indicate static stability to unsaturated overturning. An unsaturated and turbulently well-mixed convecting layer

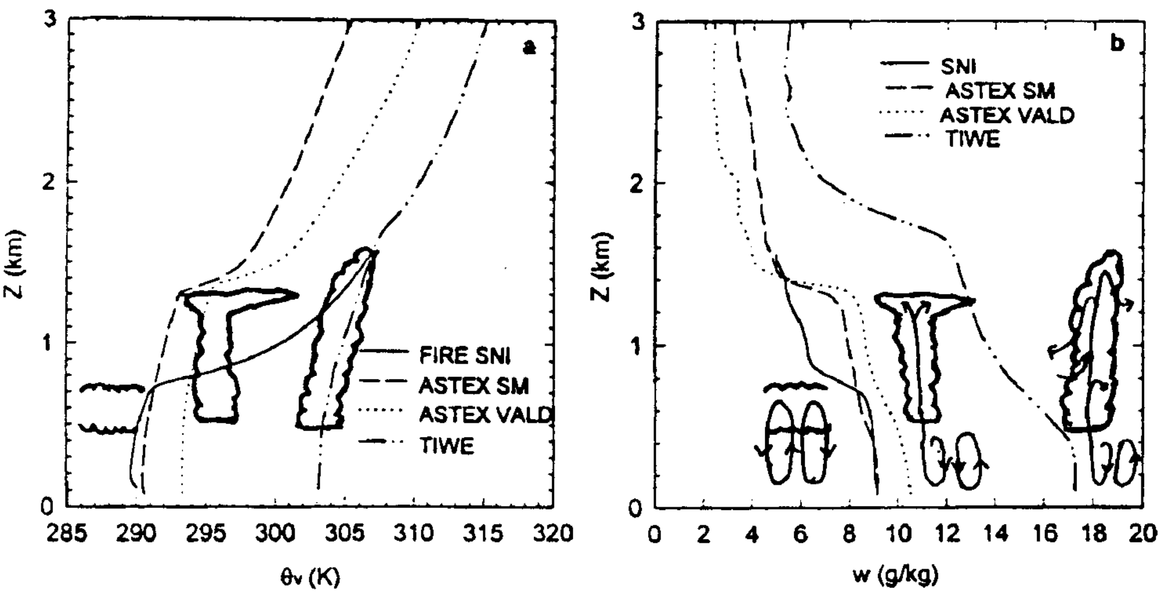



*Figure 3.* Composite soundings of (a) $\theta_v$ and (b) $q_t$ from four CTBL experiments from Albrecht *et al.* (1995). Sketches of the typical boundary layer cloud structure observed in (left to right) FIRE (July 1987, 33 N, 120 W, SST = 289 K, Cloud Fraction = 0.83), ASTEX (June 1992, SM: 37 N, 25 W, SST = 291 K, CF = 0.67; VALD: 28 N, 24 W, SST = 294 K, CF = 0.40), ), and TIWE (December 1991, 0 N, 140 W, SST = 300 K, CF = 0.26) are overlaid. In (b), the air motions that accompany the clouds are also sketched.

will approximately follow a dry adiabat in which $\theta_v$ is constant. A turbulently well-mixed, fully saturated layer will have a moist-adiabatic temperature profile in which $\theta_v$ increases with height at 3-6 K km$^{-1}$.

The second variable plotted is the total water mixing ratio $q_t = q_v + q_l$, which is the ratio of the mass of water in both vapor ($q_v$) and liquid ($q_l$) phases to the mass of dry air. This is conserved following all adiabatic motions of an air parcel, even including phase changes, as long as water is not lost to precipitation. . In boundary layer clouds, most liquid water is in cloud droplets approximately 10 microns in diameter, which fall negligibly slowly relative to the air motions, so to a first approximation many boundary layer clouds can be regarded as nonprecipitating, i. e., all condensed water remains with the air parcel in which it condensed. The total water mixing ratio is also 'linearly mixing', i. e. the $q_t$ of a mixture of two air masses will be the mass-weighted average of the $q_t$'s of the two component air masses. Hence, the vertical structure of $q_t$ carries information about the mixing history of air in different parts of the boundary layer.

In all four locations, the CTBL is capped by an inversion in which $\theta$ increases in a nearly steplike fashion, and above which the air is much drier. The boundary layer above the coldest water (SNI, a composite of sound-

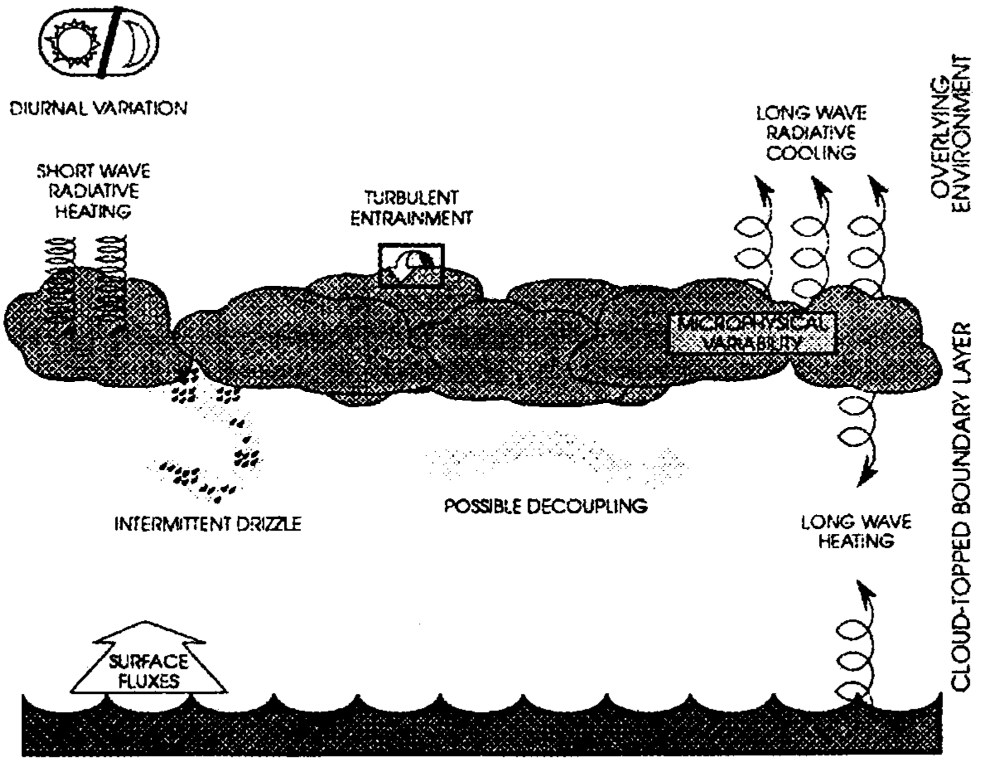
overturning. An unsaturated and turbulently well-mixed convecting layer ings taken on San Nicolas Island just off the coast of southern California during the First ISCCP Regional Experiment) has the lowest and strongest inversion and the simplest internal structure. The SNI $q_t$ profile is nearly uniform with height. In individual profiles used to build up this composite, $\theta_v$ is dry-adiabatic below cloud base and moist-adiabatic above cloud base (this is somewhat smeared out by the compositing method). This is consistent with a turbulently well-mixed CTBL.

The CTBLs above warmer water are deeper and have lower cloud amount. Observations show that the intermediate regimes (ASTEX SM and VALD, composited from soundings taken at Santa Maria Island and the R/V Valdivia (VALD) in the central north Atlantic Ocean during the Atlantic Stratocumulus Transition Experiment) are associated with cumulus clouds rising into a thin and patchy stratus layer just beneath the inversion, while the warmest CTBL (TIWE, composited from ship-based soundings taken in the eastern equatorial Pacific Ocean during the Tropical Instability Wave Experiment) supports only trade cumulus clouds. The ASTEX and TIWE CTBLs are well-mixed only below the cumulus cloud base 400-500 m above the sea surface. In the next 100-200 m (around this cloud base) there is a 'transition layer' in which moisture decreases with height and individual temperature profiles often show a thin layer of static stability. Within the cumulus cloud layer, there is a continued decrease of $q_t$ and increase of $\theta_v$ with height which is small in the intermediate profiles but considerable (but still less than moist-adiabatic) in the TIWE trade cumulus profile.

We conclude that the phase change of water has considerably complicated the dynamics of these deeper CTBLs such that neither dry nor moist convective adjustment gives a reasonable approximation to their vertical structure. It is also at first sight puzzling that the shallowest boundary layers, with the shallowest clouds, should tend to have the highest cloud cover. To probe these enigmas, we must develop a better understanding of CTBL dynamics.

## 3. Convective dynamics of CTBLs

The turbulence in CTBLs is more often than not convective, driven by radiative cooling of air in the upper part of the boundary layer and heat fluxes from the surface. We start by examining the maintenance of a shallow, well mixed subtropical stratus-capped boundary layer such as the SNI case above. Figure 4 shows the physical processes that control the evolution of such a CTBL. The most important driving mechanism for convection is infrared radiative cooling at the cloud top. The air within 50 m of the cloud top is rapidly cooled by blackbody radiation, as it emits considerably more infrared radiation than it absorbs from downwelling radiation emitted by




'igure 4. Physical processes that affect a subtropical cloud-topped mixed layer, from iems (1991)

vater vapor in the overlying atmosphere. This cooled air sinks in convecive cells that extend down to the ocean surface, where it is moistened. t is usually slightly warmed by surface heat fluxes as well. In subtropical oundary layers these are typically 10% or less of the radiative cooling that s the principal thermal driving for the convection. In mid-latitudes, parcicularly when colder air flows out over warmer water, surface fluxes can ecome dominant. When the moistened air rises, it condenses to form the loud.

The convective downdrafts and updraft speeds are around 1 m $s^{-1}$. Near he cloud-top, turbulent entrainment by the convective eddies brings some varm, dry above cloud air into the convective circulation. This would slowly leepen the boundary layer, but these CTBLs are generally found in regions f large-scale subsidence. The subsidence can counteract the entrainment leepening to permit a nearly steady state convecting layer to form in which urface moistening balances entrainment drying, radiative cooling balances ntrainment and surface warming, and entrainment balances large-scale ubsidence.

Lilly (1968), Schubert *et al.* (1979) and many others have used mixed ayer models to explore well-mixed CTBLs. These models assume that thernodynamic variables which are conserved in adiabatic motions with phase hange will be uniform throughout the mixed layer up to the inversion height $z_i$. One such variable is $q_t$. A second is the 'liquid water' potential temperature $\theta_l = \theta - cq_l$, where c = 2500 K is a thermodynamic constant that reflects the increase of $\theta$ by latent heating due to condensation of the liquid water $q_l$ in an air parcel. The evolution of the mixed layer is determined by equations for the rates of change of $\theta_l$, $q_t$ and $z_i$. $\theta_l$ and $q_t$ change due to surface fluxes, radiative cooling and entrainment. $z_i$ evolves due to specified large scale subsidence and entrainment.

A key part of a mixed layer model is an 'entrainment closure' that determines the entrainment rate $w_e$ in terms of known quantities, usually considering the budget of turbulent kinetic energy. The appropriate entrainment closure for a CTBL is controversial. In dry boundary layers heated from below, considerable evidence from observations, laboratory experiments and computer modelling shows that the entrainment rate is such that there is a downward entrainment flux of warm air from above into the boundary layer that is 20% as large as the rate of surface heating (Stull 1976). In CTBLs, the entrainment rates are small and very difficult to measure, and the above-cloud air can evaporate cloud droplets if it mixes with CTBL air, reducing the buoyancy of mixtures above the cloud base. A variety of disparate approaches have been proposed, all of which generalize the dry, surface heated case in different ways (see chapter on 'Entrainment, Detrainment and Mixing'). Stage and Businger (1981) and Nicholls and Turton (1986) have compared these closures against the very limited observational data. Nicholls and Turton's closure, which is perhaps the most appealing closure that is consistent with the data, takes the form

$$\frac{w_e}{w_*} = \frac{A}{Ri}. \tag{1}$$

Here the 'entrainment efficiency' $A$ is a parameter which increases rapidly with the maximum amount of evaporative cooling that can be produced by the mixing of cloudy and above-cloud air. For typical subtropical CTBLs, $A \approx 2$, about ten times as large as for dry surface-heated boundary layers. $w_*$ is a convective velocity scale given by

$$w_*^3 = 2.5 \int_0^H (g/\theta_0)\overline{w'\theta'}dz, \tag{2}$$

where $g$ is gravity, $\theta_0$ is a reference potential temperature, and $(g/\theta_0)\overline{w'\theta_v'}$ is the vertical buoyancy flux, which is proportional to the vertical flux of virtual potential temperature. $w_*$ can be determined from the mixed layer parameters using parameterizations of the radiative cooling rate, the surface fluxes, and $w_e$ itself. Ri is the interfacial Richardson number

$$Ri = \frac{g z_i \Delta\theta_{vi}/\theta_0}{w_*^2}, \tag{3}$$

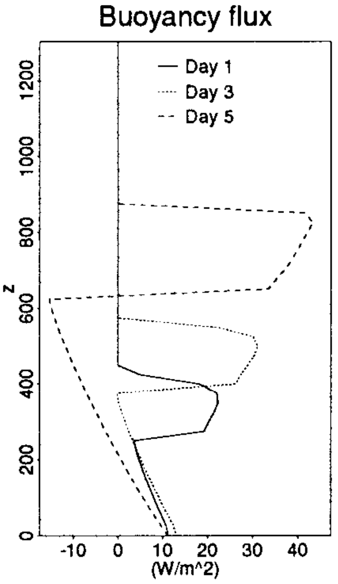



*Figure 5.* The evolution of the buoyancy flux profile in a mixed layer model of a subtropical CTBL (Bretherton and Wyant 1997) under which sea-surface temperature is raised at 1.5 K day$^{-1}$ from 285 K at Day 0 to 294 K at Day 6, while other external parameters (wind speed, mean subsidence, above-inversion temperature, mixing ratio, and downwelling solar and infrared radiation) are held constant at typical subtropical values. Numerical simulations and observations of CTBLs suggest that decoupling (i.e. the breakdown of well-mixedness in the CTBL) occurs when considerable negative buoyancy fluxes develop below cloud base, which occurs between days 3 and 5.

where $\Delta\theta_{vi}$ is the virtual potential temperature jump across the inversion. In this closure, the entrainment rate increases with the cube of the large eddy convective velocity, and decreases in proportion to the stability of the inversion. The least convincing part of the closure is the very strong dependence of $A$ on evaporative cooling, which is not well understood theoretically and is a fit to only six data points which have large uncertainties.

Lilly (1968) and others found that convective mixed layer models make good predictions of the depth, cloud thickness, and turbulence levels observed in shallow stratus-capped CTBLs off the coast of California. However, problems arose when researchers tried to apply these models on a more global scale. Wakefield and Schubert (1981) pioneered a Lagrangian approach to mixed layer evolution, treating the boundary layer air as a column moving in the wind from one location to another, thereby experiencing changing boundary conditions. In the summer, air columns move in the trade winds from locations slightly off the California coast to Hawaii in roughly a week, and the cloud is observed to change from a stratus layer to isolated cumulus clouds. The mixed layer model predicts a continuous thickening of the stratus layer instead.

Bretherton and Wyant (1997) resolved this inconsistency by showing that the well-mixedness of the convective layer breaks down as the CTBL

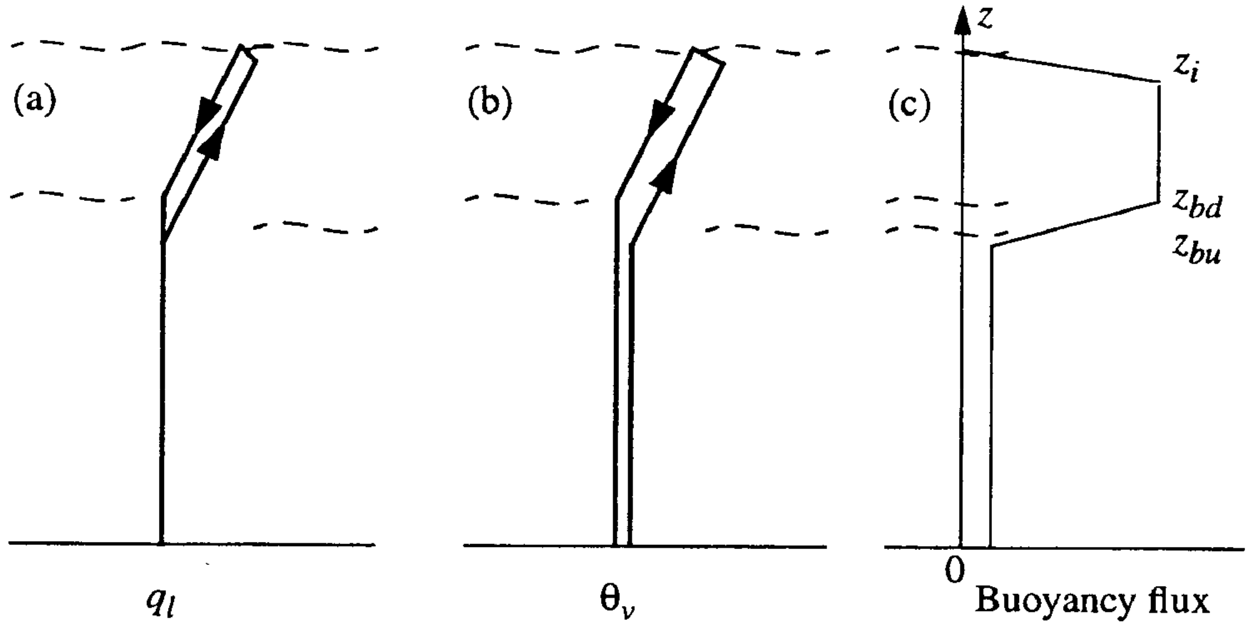



*Figure 6.* Air parcel properties and buoyancy fluxes in an convective eddy in a well-mixed CTBL. Dashed lines indicate heights of updraft condensation level $z_{bu}$, downdraft condensation level $z_{bd}$ and inversion $z_i$. Typical profiles of (a) $q_l$ and (b) $\theta_v$ for air parcels cycling through updrafts and downdrafts are shown, along with (c) the resulting buoyancy flux profile.

warms and deepens. Figure 5 shows the profile of buoyancy fluxes in a mixed layer model at three stages in the evolution of a mixed layer moving from colder to warmer sea-surface temperature. As the mixed layer deepens, it develops an expanding region of negative buoyancy fluxes below cloud base. While the mixed layer model is forced to keep the CTBL well-mixed through this process, the negative buoyancy fluxes (less dense air being forced down) tend to damp out convective motions in this region and cause 'decoupling' of the turbulent circulations into separate layers near the surface and within the cloud. Typically this occurs by the time the CTBL is 1 km deep; most convective CTBLs more than a few hundred km away from a coast (i. e. the majority of the ocean areas covered by boundary layer cloud) are probably decoupled. The formation of decoupling in a mixed layer model depends on the type of entrainment closure. Most entrainment closures prior to (1) did not permit the formation of decoupling, and could not explain why well mixed coastal stratus layers are observed to become less well mixed as they advect offshore and deepen.

Among convective systems, the CTBL is uniquely vulnerable to decoupling because of the unusual profile of buoyancy flux that is created by the phase change of water (Figure 6). Updrafts are slightly moister and have a lower condensation level than downdrafts (which have been diluted by slight mixing with entrained dry above-cloud air). Even if an updraft and downdraft have nearly the same temperature below the cloud, the updraft will have a higher temperature in the cloud, because it has more condensed liquid water, which has released more latent heat. This effect causes the buoyancy flux to rise considerably in the cloud. In fact, in subtropical cloud-topped mixed layers, the convection is typically driven mainly from within the cloud, with comparatively small buoyancy fluxes below cloud base. As the water warms, the updrafts are moistened more and more, and the difference between updraft and downdraft condensation levels increases until decoupling occurs.

The dynamics that follow decoupling have been documented in observations of deepening, warming CTBLs from the subtropical North Atlantic (Bretherton and Pincus 1995; Bretherton *et al.* 1995), and eddy-resolving numerical models of boundary layer air columns (Krueger *et al.* 1995a, b; Wyant *et al.* 1997). As the cloud base of the updrafts and downdrafts gradually separates, the updrafts become more intense and widely spaced, resembling small cumulus clouds rising into an overlying cloud layer, while the downdraft areas below the downdraft cloud base broaden.

When the inversion reaches 1500 m, the stratus layer below the inversion has started to break into broad, thin patches. It is sustained by intermittent injection of liquid water by the cumulus updrafts rising from the moister subcloud layer. These updrafts rise until they encounter the inversion, where they become negatively buoyant and detrain their air.

Figure 7 shows an idealization of air parcel circuits in a decoupled boundary layer. Most of the convective updrafts in the subcloud layer do not have sufficient inertia and buoyancy to penetrate through the weak stable layer near cloud base and form cumulus cloud updrafts; this is indicated by separating the branch of the circulation that goes up into the cumulus cloud layer from the subcloud layer circuit. The air outside the cumuli is slowly subsiding (indicated by downward arrows in the circuits of $q_t$ and $\theta_v$) and is considerably drier than the cumulus updrafts (as shown in the observational composite of Figure 3b). This contributes to drying of the cumulus updraftsby lateral entrainment, shown be the decrease of updraft $q_t$ with height. Some moist air is detrained by the cumuli, so the subsiding air moistens slightly as it descends. Penetrative entrainment by cumuli and some radiatively driven turbulence within the stratocumulus layer dry out the updraft air before it begins to subside. When the subsiding air reaches the cumulus cloud base, it is entrained into the much moister subcloud layer.

Within the subcloud layer, the circuit of $\theta_v$ shows slight radiative cooling air both as air ascends in the updraft and descents in the downdraft, with slight warming by surface buoyancy fluxes and by entrainment of warmer air

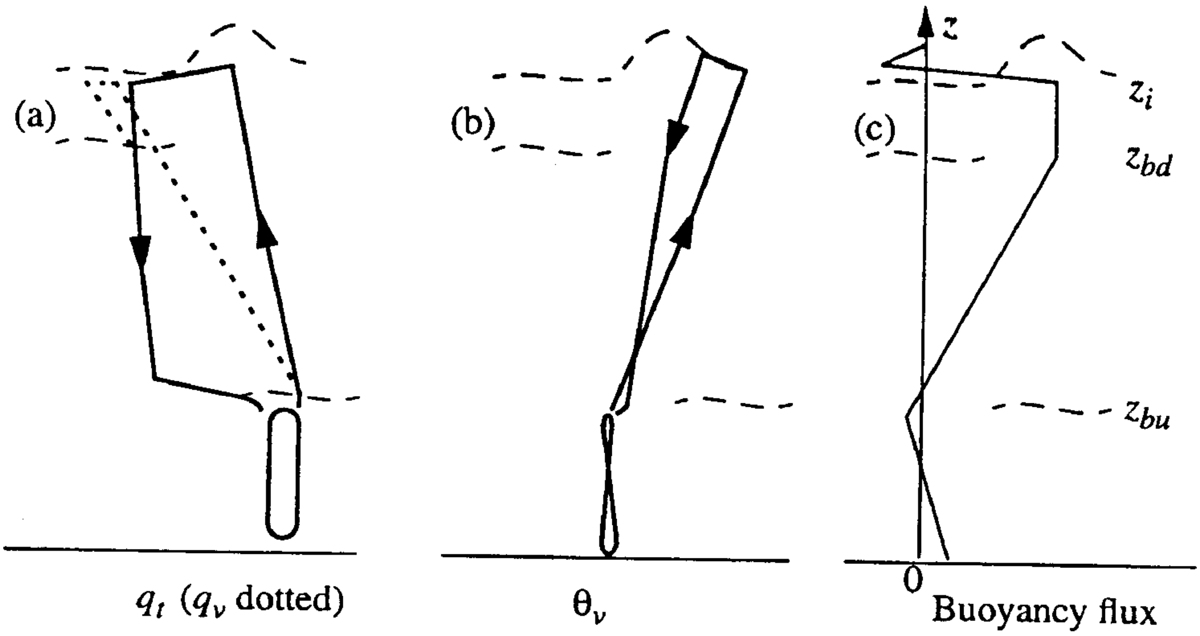



*Figure 7.* Air parcel paths and buoyancy fluxes in an convective eddy in a decoupled CTBL with cumulus rising into stratocumulus. Dashed lines indicate heights of cumulus base (updraft condensation level) $z_{bu}$, stratocumulus cloud base (downdraft condensation level) $z_{bd}$ and inversion $z_i$. Typical profiles of (a) $q_t$ amd $q_v$ and (b) $\theta_v$ for air parcels cycling through updrafts and downdrafts are shown, along with (c) the resulting buoyancy flux profile. Note that all but a few percent of subcloud layer eddies (closed circuits) do not form cumuli. See text for more discussion.

at the top of the subcloud layer. In the cumulus updrafts, there is warming due to condensation, partly counteracted by evaporative cooling due to entrainment of the ambient subsaturated air. Penetrative entrainment where the cumulus updraft encounters the trade inversion and entrains overlying above-inversion air causes more evaporative cooling, supplemented by cloud top radiative cooling as the detrained air spreads out as a patch of stratus surrounding the cumulus cloud. The subsiding air cools rapidly by evaporation as it sinks to the base of the stratus layer, then much more slowly below, due to longwave radiative cooling (see below). As in the well-mixed boundary layer, the buoyancy fluxes (consistent with the updraft-downdraft differences of $\theta_v$) are largest in the upper part of the boundary layer and are minimum at the top of the subcloud layer, but the buoyancy flux profile is somewhat more complex than in the well-mixed case.

The stratification in the cumulus layer, as shown in Figures 3a and 7b, is quite weak. This is due to the balance between convection and radiation. Extensive cloud at the inversion causes considerable infrared cooling at the inversion, but provides a greenhouse effect that makes cooling beneath the inversion very weak. Thus the air that subsides around the cumulus clouds

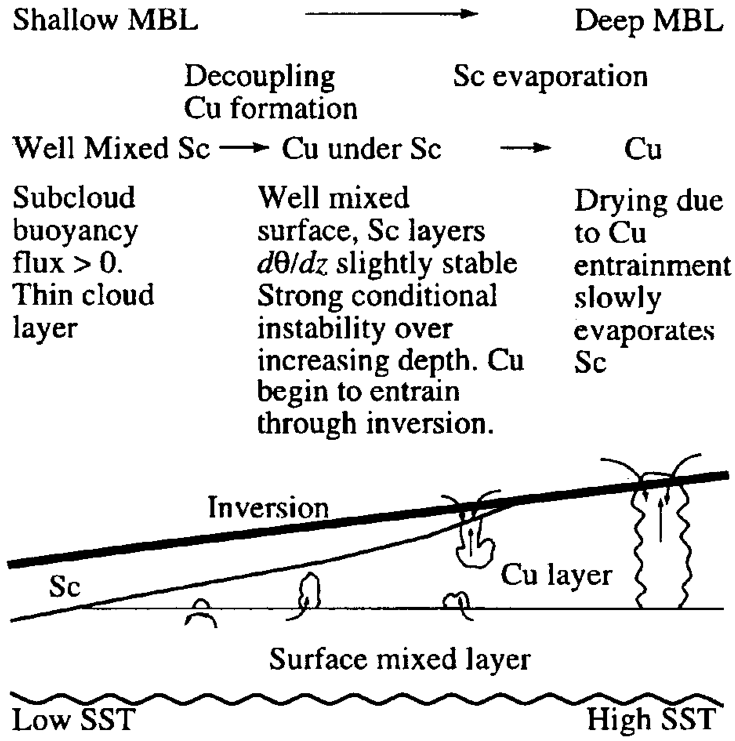



*Figure 8.* A conceptual model of the entire transition from subtropical stratus to cumulus capped CTBLs, from Wyant et al. (1997)

experiences only slight radiative cooling as it sinks through the cumulus layer, producing a very weak stratification around the cumulus clouds.

The weak stratification allows cloudy air parcels to accelerate quite rapidly as they ascend through this layer. Hence cumulus clouds overshoot their equilibrium level of neutral buoyancy at the inversion. Wyant *et al.* (1997) present evidence from their numerical model that this overshoot causes the ultimate breakup of the upper cloud layer, leaving just a cumulus cloud field as is observed over the warmer subtropical waters. The overshooting cumuli entrain some of the overlying drier, warmer air that they penetrate into. This air mixes with the cloudy updraft air, evaporating its liquid water. As the CTBL deepens, the cumulus updrafts have more distance to accelerate, so they can penetrate further into and entrain more from above the inversion. This dries the updraft air before it is detrained below and into the inversion, so that this air no longer supports a stratus cloud layer. A conceptual model of the entire transition from subtropical stratus to cumulus capped CTBLs is presented in Figure 8.

## 4. Further observations and conclusions

Many other processes affect the evolution of marine CTBLs. While boundary layer clouds typically do not precipitate heavily, they can precipitate enough to affect both the water and energy balance of the boundary layer. Precipitation processes are tightly coupled to the cloud 'microphysics', i. e. the distribution of droplet sizes within the cloud. Droplets about 5-10 microns in radius condense on small submicron diameter aerosol particles, then coalesce to form larger precipitation-size droplets 50-2000 microns in radius. The less condensation nucleii there are, the larger the initial condensed droplets are and the more readily a few of them can grow to precipitation size. In pristine marine airmasses, there are typically 50 condensation nucleii per cubic centimeter. Under these conditions, stratus clouds more than 200 m thick can drizzle. This promotes decoupling by causing water to evaporate much lower than it condenses, which causes latent heat release in the cloud and cooling below the cloud, a distribution unfavorable to convection. Cumulus clouds as little as 1 km deep in pristine airmasses can also produce showers. This depletes the cloud liquid water in cumulus updrafts so they do not detrain as much liquid water, which can considerable reduce the stratus cloud cover in the intermediate decoupled regime. In airmasses that have flowed off of polluted continents, condensation nucleus concentrations can be as much as 500 $cm^{-3}$, which almost entirely suppresses precipitation in boundary layer clouds.

If the liquid water in the clouds is subdivided into many small droplets instead of a few larger ones, this also increases the surface area of the droplets and hence their effectiveness in scattering solar radiation. While this appears to have little direct effect on the CTBL dynamics, it can considerably increase the cloud albedo. A dramatic demonstration of this is seen in Figure 9. This figure shows a satellite image of 'ship tracks', lines of brightening several hundred km long and 5-10 km wide often visible in shallow marine stratus cloud layers. These have been traced to aerosols in stack effluent that act as condensation nucleii, increasing cloud droplet concentrations by 10-500% above background values (Radke *et al.* 1987). Ship tracks are much more rarely detected in deeper boundary layers, in which decoupling makes the mixing of effluent near the surface into the main cloud layer much more intermittent and patchy. The concomitant reduction in precipitation in polluted boundary layer clouds may also increase their liquid water content and their cloud cover, further increasing their areal-average albedo (Albrecht 1989). Anthropogenic sources of cloud condensation nucleii feedbacks on boundary layer clouds and their convective dynamics may be helping to significantly increase the typical albedo of CTBLs over the oceans and counteract greenhouse gas induced global

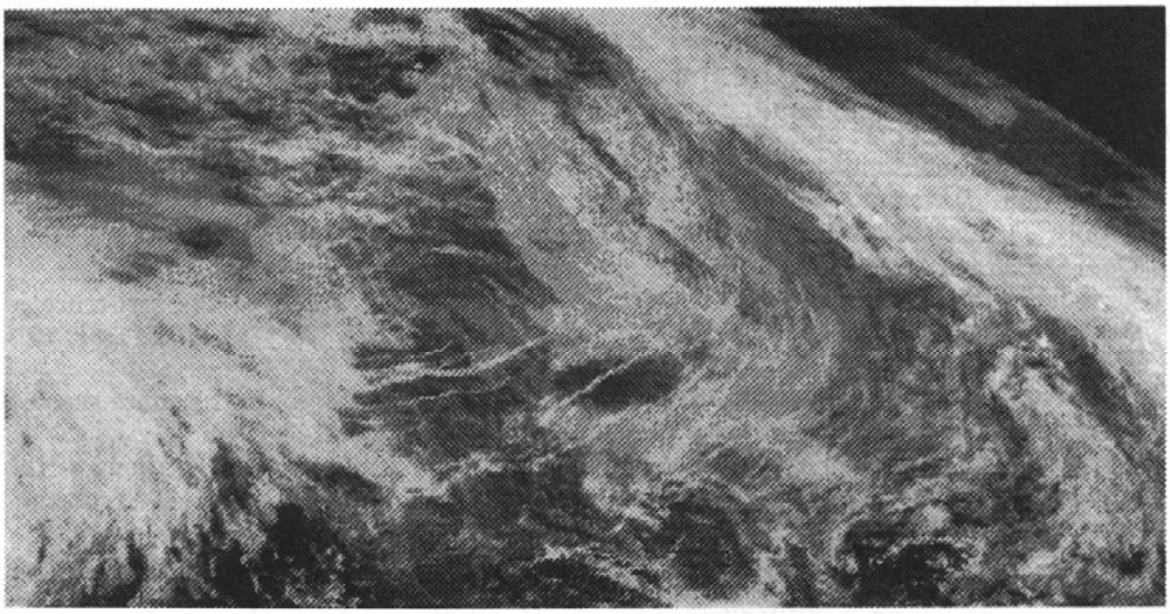


*Figure 9.* **Geostationary satellite image (in visible light) of ship tracks in stratus clouds over the summertime northeast Pacific Ocean**

warming of our current climate (Boucher and Lohmann 1995).

Boundary layer convection is surprisingly different from even deep moist convection in the atmosphere, let alone laboratory analogues. Phase change, precipitation, and the interaction of clouds and radiation considerably affect the dynamics of the convection, the vertical thermodynamic structure of the convecting layer, and the degree of horizontal homogeneity. However, numerical models and observations have been used with considerable success to understand the most important feedbacks.

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