Buoyancy reversal, decoupling and the transition from stratocumulus to shallow cumulus topped marine boundary layers

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Abstract The transition in a marine boundary layer (MBL) from stratocumulus topped to shallow cumulus topped is investigated by using a large eddy simulation (LES) model. The experiments performed aim to examine the influence on the transition of (1) the probability of buoyancy reversal at the MBL top (i.e. situations in which the mixture of two air parcels becomes denser than either of the original parcels due to phase change or other non-linear processes involved in the mixing), and (2) the degree of decoupling in the MBL (i.e. the strength of a shallow stably stratified layer near cloud base). Our results suggest that a stratocumulus-topped MBL is most likely to transit to a cumulus-topped one when (1) there exists high probability of buoyancy reversal at the MBL top, and (2) the MBL is decoupled due to large surface evaporation. We argue that a parameterization that includes representation of those two effects combined has the potential to provide a simple way of predicting the MBL transition in climate models.

Keywords Stratocumulus · Trade wind cumulus · Boundary layer · Subtropical oceans · Large eddy simulation · Climate modeling · Transition · Buoyancy reversal · Decoupling

1 Introduction

The transition from stratocumulus to trade-wind cumulus topped marine boundary layer (SMBL to TCMBL) is an

important climate feature in regions of the Pacific and Atlantic oceans under the descending branch of the Hadley-Walker circulation. Figure 1 is a schematic of this feature based on Arakawa (1975). Climatologically, near the subtropical American (African) coast the MBL is usually shallow and topped by a thin and fairly uniform layer of stratocumulus. Moving west away from the coast, clouds in the MBL become more cumulus-like (“cumulus-into-stratus”) as subsidence weakens and sea surface temperature (SST) increases. At some point the TCMBL regime becomes dominant. This TCMBL regime extends over a large region until the deep convection regime in the convergence zones. The TCMBL regime is characterized by puffy small clouds (low cloud fraction) and very active vertical transport of moisture and heat in the lower part of the troposphere. Atmospheric general circulation models (AGCMs) have great difficulties with the successful simulation of the SMBL and TCMBL regimes, and particularly the transition between them. Such difficulties are due in large part to our lack of understanding of the various MBL processes involved in the transition, which has hindered advances towards better parameterizations.

Conceptual models resulting from decades of research on the SMBL and TCMBL provided frameworks to build boundary layer parameterizations for climate models (e.g., Suarez et al. 1983; Konor et al. 2009; Neggers et al. 2006; Bretherton et al. 2004; Lock et al. 2000). The basic features in a conceptual model of SMBL include (a) well-mixed structure and (b) cloud top radiative cooling as the dominant buoyancy source. The corresponding features in a conceptual model of TCMBL are (a) distinct cloud and sub-cloud layers and (b) a conditionally unstable cloud layer driven by latent heat release in cumulus updrafts rising from the well-mixed sub-cloud layer. Within those conceptual frameworks, the main obstacle for the

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Fig. 1 A schematic showing the evolution of cloud topped MBL regimes along the trade wind track in the central to eastern South Pacific. The two vertical dashed lines broadly divide the path into three different regimes: (1) well-mixed SMBL under strong subsidence near the eastern ocean coast, (2) deep cumulus convection

regime near the South Pacific Convergence Zone, and (3) TCMBL regime in between. During the transition from SMBL to TCMBL, the so-called cumulus-into-stratus situation (as depicted above) is often observed to occur

simulation of the SMBL-to-TCMBL transition lies in the difficulty to account accurately for the active roles played by both cloud-top radiative forcing, cloud-top entrainment and surface controlled cumulus updrafts in driving non-local turbulent mixing during the transition. To overcome such obstacles, one could either build a unified parameterization capable of describing both MBL regimes or develop a treatment for the transition between the two. In this paper, we concern ourselves with the second possibility. In this context, we investigate the existence of parameters that enable us to predict when the SMBL regime transits to the TCMBL regime.

Lagrangian studies of the MBL along trade wind tracks focused specifically on the impact of various turbulent, radiative and microphysical processes involved in the SMBL-to-TCMBL transition (e.g., Krueger et al. (1995). Wyant et al. (1997) and Bretherton and Wyant (1997; hereafter collectively referred to as WB97) proposed a conceptual model of the transition. The key process in their model is the development, in the cloud layer, of cumulus updrafts, driven by increased surface evaporation. This development leads to deepening of the cloud layer even when the stratocumulus clouds near the MBL top are being destroyed. In addition, the cloud layer decouples from the subcloud layer. WB97 suggested that whether the MBL is decoupled or not is a simple way to judge which one of the two conceptual frameworks mentioned above fits better. It is possible, however, that stratocumulus clouds, after decoupling, still play a secondary but active role in their own demise (e.g., Stevens et al. 2001).

From a rather different perspective, instead of emphasizing the importance of surface forcing during the transition, Randall (1987) considered only the effects of the

changes in conditions across the inversion in his mass flux model of the MBL. He postulated that when the buoyancy reversal criterion (Randall 1980; Deardorff 1980) is violated, only a cumulus regime with small updraft fraction is stable to a process referred to as the cloud top entrainment instability (CTEI, see Deardorff 1980). According to Randall's (1987) results, buoyancy reversal by itself dictates the transition from a well-mixed SMBL to a TCMBL. Studies based on observational data and LES simulations, however, have suggested that overcast SMBLs can exist even when the buoyancy reversal criterion is violated (e.g., Moeng 2000; Kuo and Schubert 1988).

We next explore whether the two perspectives of the transition described above, one emphasizing control from the surface and the other control from the free atmosphere above the inversion, could be reconciled to provide a better criterion for the transition than that obtained from either perspective alone. We start by looking at the two key concepts—decoupling and buoyancy reversal—in greater detail.

1.1 Decoupling

Decoupling refers to the situation in which a shallow stably stratified zone exists near the cloud base in a MBL. This stable zone in the mean profiles roughly coincides with local minima in turbulent kinetic energy and vertical buoyancy flux. Symptoms of decoupling include increased cloud horizontal inhomogeneity at the mean cloud base level, and the separation of the mean lift condensation level from the mean cloud base level (e.g. Zuidema et al. 2009). In the SMBL, decoupling could happen for various reasons not necessarily independent of each other: (1) SST cooling

(e.g., Paluch et al. 1999; Pyatt et al. 2005); (2) development of cumulus updrafts in the cloud layer due to increased surface evaporation and/or decreased Bowen ratio, the ratio between surface sensible and latent heat flux (e.g., Krueger et al. 1995) and (3) destruction of the overlying stratocumulus layer for reasons other than the development of cumulus updrafts (e.g., drizzle and short-wave radiative warming, see WB97). A quantitative definition of the degree of decoupling was given by the observational studies of Park et al. (2004) and Wood and Bretherton (2004),

$$\alpha_q = \frac{q_t(z = z_i^-) - q_t(z = 0)}{q_t(z = z_i^+) - q_t(z = 0)}, \quad (1)$$

$$\alpha_\theta = \frac{\theta_l(z = z_i^-) - \theta_l(z = 0)}{\theta_l(z = z_i^+) - \theta_l(z = 0)},$$

where q_t and θ_l are total water mixing ratio and liquid water potential temperature respectively; z_i^+ and z_i^- are vertical levels immediately above and below the cloud top level (z_i). $\alpha_q(\alpha_\theta)$ measures the moisture decrease (thermal stratification) across the cloud layer against that across the inversion. α_q and α_θ are not equal due to the diabatic cooling effect of cloud. Wood and Bretherton (2004) estimated that α_q and α_θ over subtropical Eastern Pacific range from ~ 0 to ~ 0.4 . $\alpha_q(\alpha_\theta) > 0$ is required for decoupling. The exact critical value, however, is uncertain.

Only decoupling due to reason (2) above, i.e., the development of cumulus updrafts, is relevant to our discussion here and we concentrate on this aspect in the remainder of the paper. In this case, decoupling happens because descending air parcels become positively buoyant and liquid-water-free before they reach the updraft condensation level. This asymmetry between ascending and descending air is caused by the increasingly important contribution of latent heat release in the updrafts to in-cloud buoyancy generation compared to the contribution by stratocumulus-top radiative cooling, which, in turn, is caused by the increase in the surface latent heat flux and deepening of the cloud layer. In other words, the cloud layer becomes more cumulus-like. WB97 used the degree of decoupling as a parameter to diagnose the breakdown of the mixed layer assumption along a Lagrangian trade wind track with increasing SST. They also proposed a definition of decoupling parameter, r , as the following:

$$r = \frac{\int_{z < z_i^-, \overline{w'b'} < 0} \overline{w'b'} dz}{\int_{\overline{w'b'} \geq 0} \overline{w'b'} dz}, \quad (2)$$

where $\overline{w'b'}$ is the vertical turbulent buoyancy flux, z_i^- is the level just below the inversion. r is the ratio between the negative buoyancy generation, near the stable zone at cloud base, and the positive buoyancy generation elsewhere in the SML. When $r > 0.15$, the mixed layer assumption is

no longer valid, according to WB97. Like many other empirical parameters, the exact critical value of r is subject to debate (for example, Stevens (2000) argued that for $r > 0.0$, instead of 0.15, decoupling already exists). The breakdown of the mixed layer assumption is not necessarily followed by swift transition to TCMBL and destruction of the stratocumulus layer. The decoupling parameters, therefore, only serve to diagnose the relative importance of the stratocumulus clouds compared to the cumulus clouds. The actual control of the transition comes from either the surface fluxes, which drive the cumulus updrafts, or the free atmosphere above the inversion.

1.2 Buoyancy reversal

Buoyancy reversal refers to the situation in which the mixture of two air parcels becomes denser than either of the original parcels due to phase change or other nonlinear processes involved in the mixing. Such a situation can occur in the entrainment mixing between the dry and warm air above the SML and the cloudy and cool air in the SML. The criterion for buoyancy reversal to occur during the entrainment mixing process has the following general form,

$$k = \frac{(\Delta s_v)_{crit}}{\Delta s_v} > 1, \quad (3)$$

In Eq. (3), Δs_v is the buoyancy excess of the above-SML air in reference to the in-SML air near the layer's top. The expression can also be interpreted as a measure of the thermal stability of the inversion. The numerator $(\Delta s_v)_{crit}$ represents an estimate of buoyancy reduction that would be produced by the evaporation of cloud water when above-SML air is mixed with in-SML air.

The particular form of $(\Delta s_v)_{crit}$ varies among different works. Randall (1980) defined it in terms of the above-SML air saturation deficit $(\Delta s_v)_{crit} \propto (q_s^+ - q^+)$, where q_s^+ and q^+ are the saturation mixing ratio and mixing ratio immediately above the SML, respectively. Adopting this definition leads to another very commonly used form of the threshold for buoyancy reversal, as given by Randall (1980) and Deardorff (1980),

$$\kappa = \frac{c_p \Delta \theta_e}{L \Delta q_t} = 1 + \frac{c_p \Delta \theta_l}{L \Delta q_t} > 0.23, \quad (4)$$

where Δq_t , $\Delta \theta_l$ and $\Delta \theta_e$ are the jumps of total water mixing ratio, liquid water potential temperature and equivalent potential temperature across the SML top respectively. Further, c_p is specific heat of air at constant pressure and L is latent heat of water evaporation. When $\kappa > 0.23$, $k > 1$ (for the definition of $(\Delta s_v)_{crit}$ by Randall (1980)). Siems et al. (1990) proposed another definition of $(\Delta s_v)_{crit}$ taking into account the air properties (e.g., liquid water amount)

below the inversion. According to Siems et al. (1990), the buoyancy reversal criterion is,

$$D = -\frac{b^*}{b^+} = -\frac{s_v^* - s_v^-}{s_v^+ - s_v^-} = \frac{s_v^+ - s_v^*}{s_v^+ - s_v^-} - 1$$

$$= \frac{(\Delta s_v)_{crit}}{\Delta s_v} - 1 > 0, \quad (5)$$

The buoyancy reversal criterion in Eq. (5) converges to that given by Eq. (3) with $(\Delta s_v)_{crit} = s_v^+ - s_v^*$, where s_v^* is the minimum virtual static energy the mixture of the above and below inversion air could have and s_v^- and s_v^+ are the virtual static energy of the air just below and above inversion respectively. s_v^* depends on the air properties both in and above the MBL.

Parameters such as k , κ and D are just measures of how likely buoyancy reversal would happen. In this regard, buoyancy reversal criteria based on these parameters only state that a certain portion of all possible mixtures between above and below inversion air would be negatively buoyant compared to the below inversion air. Different definitions of $(\Delta s_v)_{crit}$ result from different ways of specifying that portion out of all possible mixtures. [For more discussions on the definitions of and relations among k , κ , D and other such parameters, the readers are referred to Yamaguchi and Randall (2008) and Shao et al. (1997).] Whether and for how long these negatively buoyant mixtures would be realized during entrainment mixing is not specified. The dynamic consequence of the possible formation of these negatively buoyant mixtures is not known either.

Several studies have addressed the role played by buoyancy reversal in the breakup of stratocumulus. Here we give only a brief summary. Randall (1980) and Dardorff (1980) proposed that when $\kappa > 0.23$, cloud top entrainment instability (CTEI) would occur, stratocumulus cloud would be destroyed and transition to TCMBL ensues (Randall 1987). A series of observational and modeling works has challenged the viability of a CTEI-based scenario of stratocumulus destruction. Kuo and Schubert (1988) and Moeng (2000), for example, found no explosive entrainment or fast stratocumulus breakup when $\kappa > 0.23$ in both observations and LES results. Nevertheless, some modification have been attempted. Siems et al. (1990) proposed that $D > 1.3$ is necessary for CTEI on the basis of both theoretical arguments considering the energetics of the entraining eddies near the SMBL top and results from laboratory and numerical experiments. MacVean and Mason (1990) proposed that $\kappa > 0.7$ is necessary for CTEI based on a cloud layer potential-to-kinetic energy conversion analysis. In observations and numerical simulations, however, stratocumulus frequently break up and become shallow cumulus below these revised thresholds (e.g., Fig. 1 in Kuo and Schubert 1988). More fundamentally, many authors (e.g. Moeng 2000; Lewellen and Yo 1993;

Siems et al. 1990) have pointed out that the impact of buoyancy reversal on SMBL can not be fully understood based solely on the consideration of entrainment evaporation at the cloud top. The most important additional factor is the turbulence structure inside the SMBL, i.e., whether the SMBL is well-mixed or decoupled.

1.2.1 Well-mixed SMBL

For “classical” well-mixed SMBLs, the dominant driver for MBL turbulence is cloud-top radiative cooling, so much so that smoke cloud (i.e., an approximation in which the latent heat effect of evaporation/condensation is ignored) is a reasonable approximation (Lilly 1968; Bretherton et al. 1999; Stevens and Bretherton 1999; Shao et al. 1997). In smoke cloud scenarios, there is no entrainment enhancement due to evaporative cooling. Increase in κ would only cause increase in the evaporation of cloud water due to stronger entrainment drying, which then leads to entrainment reduction rather than enhancement. Thus, for shallow and well-mixed SMBLs the entrainment enhancement by evaporative cooling may be overcome by the entrainment reduction due to cloud destruction, thus preventing fast stratocumulus destruction.

1.2.2 Decoupled SMBL

In decoupled SMBLs, the changes near the cloud top and near the surface are communicated throughout the layer by very different processes. The changes near the surface are carried into the cloud layer by cumulus updrafts directly while the changes near the cloud top are communicated downward by the less efficient downward mixing and by entrainment across the transition layer. It follows that the entrainment drying effect would be more concentrated near the cloud top than in the well-mixed SMBL. This could potentially lead to further reduction in cloud water content, which would further amplify the concentration of drying and warming at the cloud top with reduced downward mixing due to reduction in cloud-top radiative cooling. This concentrated entrainment drying is an additional mechanism for stratocumulus destruction at the expense of decoupling. In this case, then, stratocumulus destruction may be self-promoting even without enhanced entrainment by evaporative cooling (see WB97).

Our review of previous work, then, suggests four possible scenarios in terms of the transition from SMBL to TCMBL according to whether the SMBL is coupled or decoupled, and whether κ is large or small (see Table 1). Our hypothesis is that a decoupled SMBL with a high probability of buoyancy reversals [Scenario (A)] is highly likely to transit to a TCMBL. The distinction between Scenario (A) and the other scenarios in Table 1, therefore,

Table 1 Possible SMBL scenarios for the transition from a SMBL to a TCMBL

	Decoupled SMBL	Well-mixed SMBL
Large κ	(A) Stratocumulus is destroyed, and the transition is highly probable.	(B) Stratocumulus is reduced, but the transition is highly unlikely
Small κ	(C) Stratocumulus and the “controls” from the cloud-top on the underlying cumulus cloud layer is still significant.	(D) No transition occurs.

must be considered in large-scale parameterizations of MBLs. Scenario (B) is tending towards a cloud-free MBL due to cloud reduction by entrainment but will not transit to a TCMBL. Scenario (C) can be seen as the second step in the two-step conceptual model of WB97, which is in a transitional regime and the control exerted on the decoupled cloud layer by cloud-top longwave radiative cooling and cloud-top entrainment are still significant even though the cloud layer is more and more cumulus-like. Scenario (D) is the typical stable, well-mixed SMBL and is of no relevance to this study.

To test our hypothesis we use a LES model, in which all the essential elements of the SMBL, i.e., turbulence, radiation and moisture, are considered albeit with limited complexity. The distinction between decoupled and well-mixed SMBLs is set up through modifying both surface forcings and initial profiles. The buoyancy reversal parameter, κ , is altered through changes in the free atmosphere moisture while the buoyancy jump across the inversion is kept the same.

We start in Sect. 2 by introducing the LES model we use and our experiment setup. Section 3 presents the results obtained in the numerical experiments. Section 4 explores the sensitivity of our main results to changes in the LES experiment setup. Section 5 concludes and discusses the implication of our results on MBL parameterizations in GCMs.

2 Model and experiment design

The large eddy simulation (LES) model we use is the UCLA LES (Stevens et al. 1999; 2005; Stevens and Seifert 2008), of which code and documentation are available from <http://www.atmos.ucla.edu/bstevens/Research/UCLALES.htm>. The model performance has been illustrated in a series of intercomparison studies on boundary layer clouds (e.g., Stevens et al. 2001; 2005; Siebesma et al. 2003). The basic setups for the simulations we perform in this study follow those used in the GCSS boundary layer working group intercomparison studies of (1) the second Dynamics and Chemistry of Marine Stratocumulus field study (DYCOMS-II) RF01 case (Stevens et al. 2005), and (2) the Atlantic Trade Wind Experiment (ATEX) (Stevens et al. 2001). The former type of setup (referred to as the MIXED

series hereafter) is typical of shallow, well-mixed SMBLs over eastern subtropical oceans under very strong inversion. The latter type of setup (referred to as the DECOUPLED series hereafter) has high stratocumulus coverage but with a substantially deeper MBL and the cloud layer is decoupled from the sub-cloud layer. This special situation is sometimes referred to as the cumulus-into-stratus regime. For each type of setup, we carry out multiple experiments with different buoyancy reversal conditions by specifying different initial values of κ . These experiments allow for an examination of the stratocumulus cloud behavior and MBL structure in all four scenarios in Table 1. Ideally we would also like to vary the surface temperature or the surface fluxes to produce more cases with different degrees of decoupling, which would allow for more detailed investigation of the cloud regime transition along the trade wind track. Without detailed observational studies and model verifications, however, we are uncertain how well the LES performs under these setups.

2.1 Model domain and duration of simulations

The DECOUPLED series setup has a horizontal resolution of 100 m in both directions and a vertical resolution of 20 m below 1,750 m. Stretching vertical grids are used between 1,750 m and the model top at about 3,000 m. In the horizontal the domain has 96×96 grid points covering an area of about 9,600 m $\times$ 9,600 m. In the vertical the domain has 131 grid points. The MIXED series setup has a horizontal resolution of 35 m in both directions and a vertical resolution of 10 m below 925 m. Stretching vertical grids are used between 925 m and the model top at about 1,500 m. The total number of horizontal and vertical grid points is the same in both setups. The horizontal domain thus is smaller, covering an area of about 3,500 m $\times$ 3,500 m. A damping, or sponge, layer is specified near the upper boundary in both setups to prevent problems with gravity wave reflection and energy accumulation.

All the simulations are initialized with random perturbations added to the spatially uniform temperature and moisture fields and are 12-h long. The initial profiles, large-scale forcing, radiative and surface fluxes are specified in the following subsections.

2.2 Initial profiles

Figure 2 shows the initial profiles of total water mixing ratio and liquid water potential temperature for the standard MIXED and DECOUPLED cases. The standard initial profiles in the DECOUPLED series follow exactly Stevens et al. (2001). The above-cloud total water mixing ratio (denoted by q_t^+ in the remainder of the paper) is 4.5 g/kg in the standard case. To produce different values of κ , q_t^+ is varied from 1.5 to 6.5 g/kg at an interval of 0.5 g/kg to yield ten more cases (eleven in the series). Since q_t^+ is the only variable parameter among the setups of different cases, we use it as the case-identifier of different cases. The initial wind profiles and their geostrophic values are specified as in Stevens et al. (2001).

For the MIXED series, the standard initial profiles are specified in the following way:

$$\theta_l = \begin{cases} 289 & \text{when } z < z_i \\ \theta_l^+ + (z - z_i)^{1/3} & \text{when } z > z_i \end{cases} \quad (6)$$

$$q_t = \begin{cases} 9.0 & \text{when } z < z_i \\ q_t^+ & \text{when } z > z_i, \end{cases}$$

where z_i is the cloud top height, taken as 840 m here. In the rest of the paper z_i for both the DECOUPLED and the MIXED series is defined as the height of maximum θ_l gradient for the domain averaged vertical profile. The values of the above-cloud liquid water potential temperature (θ_l^+) and total water mixing ratio (q_t^+) are 297.5 K and 1.5 g/kg respectively in Stevens et al. (2005). To allow for more extreme values of κ , θ_l^+ is reduced to 292.5 K in our setup. q_t^+ then is varied from 0.5 to 4.5 g/kg at an interval of 0.5 g/kg, yielding nine cases in total. q_t^+ is also used as case-identifier in the MIXED series.

Fig. 2 Initial profiles of total water mixing ratio (g/kg) and liquid water potential temperature (K) for the standard DECOUPLED and MIXED cases

2.3 Large-scale subsidence, radiative forcing and surface fluxes

The large-scale subsidence in the DECOUPLED series is prescribed to be 6.5 mm/s at the cloud top level (z_i), and to decay linearly both downward and upward to zero at the surface and at 300m above the cloud top. For the MIXED series, a constant divergence rate of $3.75 \times 10^{-6} \text{ s}^{-1}$ is specified. No shortwave radiation fluxes are included in the simulations. In setting up our experiments, we think it is a reasonable approximation to consider the impact of shortwave radiative warming in the cloud layer on decoupling and cloud layer structure secondary to that of the surface forcing for cloud regime transitioning along the trade wind track. As in Stevens et al. (2001) we use a simplified longwave radiation flux of the following form,

$$F_{rad} = F_0 \exp \left(-\alpha \int_z^{\text{inf}} \rho l dz \right), \quad (7)$$

where $F_0 = 74 \text{ W m}^{-2}$, $\alpha = 130 \text{ m kg}^{-1}$. ρ is density and l is liquid water mixing ratio. In some of our experiments producing small cloud fraction (~ 0.2), the chance of forming cloud-free columns increases. Our longwave radiation flux calculation does not take into account clear-sky radiative cooling for the clearings in the cloud layer. This may lead to overestimation of the temperature in the cloud-free downdrafts and affect the cloud layer circulation.

The surface heat fluxes both in space and time are prescribed to be constant. In the DECOUPLED series, the surface sensible heat and latent heat fluxes are 13.3 and 132 W m^{-2} , respectively. The corresponding values in the MIXED series are 16 and 93 W m^{-2} , respectively. The effect of precipitation processes on clouds is considered separately in the sensitivity experiments described in Sect. 4.

3 Results

We first present vertical profiles of domain/time average quantities. Figure 3 shows the last 3-h mean profiles of horizontally averaged liquid water potential temperature, liquid water and total water mixing ratio for the MIXED series (upper panels) and the DECOUPLED series (lower panels). The time series of domain averaged cloud coverage, liquid water path (LWP) and cloud top height for the total duration of all runs are presented in Fig. 4. Vertical profiles of higher order statistics for $\overline{w'w'}$, grid-scale buoyancy production and longwave radiation flux are shown in Fig. 5.

In the MIXED series the mixed layer structure is maintained, even though cloud amount is considerably

Fig. 3 Vertical profiles of liquid water mixing ratio (**a** for the MIXED series, **d** for the DECOUPLED series), liquid water potential temperature (**b** for the MIXED series, **e** for the DECOUPLED series) and total water mixing ratio (**c** for the MIXED series, **f** for the

DECOUPLED series). Values correspond to averages over the model domain and last 3 hours of the simulations. The profiles are color coded according to the initial values of above-inversion total water mixing ratio

reduced in some cases. The MBL temperature (Fig. 3a) increases slightly due to less cloud cover for the cases with lower initial q_t^+ . The MBL total water mixing ratio (Fig. 3b) is also reduced for cases with lower initial q_t^+ . The liquid water amount (Fig. 3c and Fig. 4a) decreases almost linearly with initial q_t^+ . Cloud fraction (Fig. 4b) higher than 60% is maintained for all cases and a smooth decrease with initial q_t^+ is found. The cloud top height evolution in Fig. 4c indicates that the overall entrainment rate decreases with initial q_t^+ . There is no sign of abrupt CTEI or “runaway” entrainment, in agreement with other works (e.g., Yamaguchi and Randall 2008; Moeng 2000). Larger values of κ computed from the initial profiles of q_t and θ_t actually lead to smaller entrainment rates. This is inconsistent with the evaporatively enhanced entrainment mechanism proposed by Siems et al. (1990), which emphasizes the enhancement of entraining eddies due to

increased evaporative cooling with larger κ near cloud top. The evaporatively enhanced entrainment formulation in Siems et al. (1990), however, did not consider the possible role played by cloud radiative processes. Our interpretation of the discrepancy with Siems et al. (1990) is that larger κ causes larger entrainment drying with the same thermal stability across the inversion, which reduces cloud coverage and cloud top radiative cooling, and thus leads to less entrainment. There is a potential competition between the effects of evaporative drying (cloud destruction) and cooling on entrainment. In view of the uncertainty on how realistically LESs can simulate cloud top entrainment (e.g., Mellado et al. 2009), there can be questions on whether evaporatively enhanced entrainment is actually overshadowed in well-mixed SMBLs by the effects of cloud destruction on entrainment. From a more theoretical point of view, we could argue that for well-

Fig. 4 Time evolution of horizontally averaged cloud fraction (a MIXED series, d DECOUPLED series), liquid water path (b MIXED series, e DECOUPLED series) and cloud top height

(c MIXED series, f DECOUPLED series). The profiles are color coded according to the initial values of above-inversion total water mixing ratio

mixed SMBLs since cloud top radiative cooling is the dominant buoyancy source, the impact of evaporatively enhanced entrainment may be marginal compared to that of cloud destruction on entrainment. This would be

consistent with the commonly held view that stratocumulus in shallow, well-mixed SMBL can be approximated as “smoke clouds”, which are only radiatively active as reviewed in the introduction.

Fig. 5 Profiles of $\overline{w'w'}$ (a MIXED series, d DECOUPLED series), grid-scale buoyancy production (b MIXED series, e DECOUPLED series), and longwave radiation flux (c MIXED series, f DECOUPLED series). Values correspond to averages over the model domain

The $\overline{w'w'}$ profiles in Fig. 5a show a single maximum in the vertical. For cases with high cloud cover (reddish lines), this maximum is located closer to cloud top at about 700 m [$\sim 0.7z_i$ ($\sim 1,000$ m)]. In cases with lower cloud cover (bluish lines), the maximum is closer to the surface at about 450 m [$\sim 0.5z_i$ (~ 900 m)]. This difference in height is consistent with the distinction drawn between a SML and a dry boundary layer driven solely by the surface fluxes in previous studies (e.g., Moeng 1986). In cases with high cloud cover buoyancy production (Fig. 5b) near the cloud top by divergence of longwave radiation flux (Fig. 5c) is much larger than that near the surface. In cases with low cloud cover buoyancy production integrated over the cloud layer becomes considerably less than that over the sub-cloud layer.

In the DECOUPLED series, the liquid water potential temperature and total water mixing ratio profiles (Fig. 3d–e) generally show a sub-cloud layer that is well mixed, and a

and the last 3 h of the simulations. The profiles are color coded according to the initial values of above-inversion total water mixing ratio

cloud layer with slowly decreasing moisture and increasing temperature with height. The liquid water profiles (Fig. 3f) show either cumulus-into-stratus structure (reddish lines) or trade cumulus structure (bluish lines). The domain averaged liquid water amount and thickness of the stratocumulus layer varies greatly with initial q_t^+ . For initial $q_t^+ \geq 4.5$ g/kg, a stratocumulus layer at the top with cloud cover larger than 80% (see Fig. 4d) is maintained, while for lower initial q_t^+ cloud cover is less than 40%. This discrete jump in cloudiness is linked to differences in the vertical profiles of total moisture and liquid water potential temperature in Fig. 3e–f. A mixed cloud layer structure is more apparent in cases with high cloud cover (reddish lines) and large initial q_t^+ than in cases with low cloud cover (bluish lines). The discrete jump at initial $q_t^+ = 4.5$ g/kg is also seen in the time series of LWP and cloud top height (Fig. 4e–f). For initial q_t^+ less than 4.5 g/kg the LWP and cloud top height are almost indistinguishable for the different cases and considerably smaller

compared with high initial q_t^+ cases during the last 3 h of simulation. In the DECOUPLED series, the contribution to buoyancy production (Fig. 5e) by latent heat flux is apparent in the cumulus part of the cloud layer. For initial $q_t^+ > 4.5$ g/kg buoyancy generation due to radiative cooling (Fig. 5f) in the cloud layer, however, is much stronger. This is consistent with the better-mixed cloud layer temperature and moisture profiles shown in Fig. 3e–f. The downward mixing driven by radiative cooling, however, is not efficient enough to mix all the way down into the sub-cloud layer because the cloud layer in these cases is already more than 800 m deep while the total MBL depth in the MIXED series is only around 1,000 m. Decoupling is manifest for all DECOUPLED cases in the mean profiles of temperature and moisture according to the definition of Park et al. (2004) and in the buoyancy profiles according to the definition of WB97. The degree of decoupling defined by WB97, calculated from last 3-h averages, ranges from 0.065 to 0.093.

To obtain a more quantitative evaluation of the SMBL in transition we examine in Fig. 6 cloud fraction and LWP as a function of κ . All values are horizontally averaged over the entire domain and again averaged over the last three hours. The above and below inversion levels used to calculate the jump are defined as 100 m (50 m) over and under the horizontally averaged z_i for the DECOUPLED (MIXED) series, respectively. Most values of κ calculated in this way fall between 0.23 and 0.7 (with only two cases in which $\kappa < 0.23$). In the MIXED series the values of κ are largely independent of the choice of reference level; in the DECOUPLED series the values of κ are only slightly more sensitive to that choice. In the dependence of both cloud fraction and LWP with κ the MIXED series shows an almost monotonic decrease with increasing κ until the limit of reasonable values for above-inversion total water mixing ratio. Rather than transitioning into a TCMBL, the SMBL becomes similar to a dry MBL driven mainly by the surface buoyancy flux (recall Fig. 5c). Interestingly, the clouds in the MIXED series, remain stratiform and the lowest cloud fraction is around 0.6. The DECOUPLED series, in contrast, shows a clear transition from cumulus-into-stratus regime to typical trade-cumulus regime separated at around $\kappa = 0.4$, which corresponds to the case with initial $q_t^+ = 4.5$ g/kg. This marked transition is not associated with a similarly marked increase in entrainment. The growth rate of cloud top height in these cases actually decreases with κ as the stratocumulus are destroyed (see Fig. 4f). Our interpretation for the stratocumulus destruction with $\kappa > 0.4$ is as follows. In the decoupled cloud layer, the downward mixing of entrained dry air near the cloud top is carried through by eddies partly driven by cloud-top radiative cooling. With larger κ , entrainment drying would be stronger with the same entrainment efficiency. This leads to cloud reduction and weakening in the cloud-top radiative

Fig. 6 Scatterplots of κ versus cloud fraction (left) and versus liquid water path (right). κ is defined in Eq. (4). All values are averages over the last three hours of the simulations. Blue triangles represent cases in the DECOUPLED series, and red squares represent cases in the MIXED series

cooling. As a consequence the downward mixing eddies are weakened. This reduced downward mixing then tends to trap the entrainment drying effect closer to the cloud top, which leads to more effective destruction of clouds, especially of stratocumulus. Thus, a positive feedback loop forms for the destruction of stratocumulus, with the cloud layer becoming more and more decoupled. For this feedback loop to be effective, a large value of κ is required. This positive feedback loop works only in the decoupled MBL because the impact of large κ is mostly trapped in the cloud layer due to decoupling. The cumulus updrafts rising from the subcloud layer and the cloud-top entrainment efficiency associated with them are little affected by increasing κ . Thus the total entrainment is maintained to enable the positive feedback despite the reduction in entrainment due to reduced cloud-top radiative cooling.

Figure 7 shows probability density functions of LWP calculated from three-dimensional snapshots (5 in total) taken during the last two hours of the simulations. In the MIXED series, the probability density functions become more flat and asymmetric with increasing κ (i.e., decreasing initial q_t^+), thus increasingly depart from a normal distribution. This feature of the LWP distribution in large κ cases also indicates that cloud fraction decreases while the patchiness of the stratocumulus cloud increases with increasing κ . The higher probability of buoyancy reversal, therefore, leads to local clearing of cloud water through similarly local enhancements of entrainment. The intensity and extent of the eddies involved in the clearing, however, seem to be limited since cloud clearing leads to reduction of buoyancy generation by radiative cooling in the entraining eddies. This is consistent with the arguments of Lewellen and Yoh (1993). The DECOUPLED series is again divided by the case with $q_t^+ = 4.5$ g/kg. For higher initial q_t^+ , the distribution is a normal one (as is typical of stratocumulus cloud layers) with a “fatter” tail (as is

Fig. 7 PDF distributions of liquid water path calculated from 5 snapshots (one every 30 min) during the last two hours of all simulations for the **a** MIXED series, **b** DECOUPLED series. The lines are color coded according to the initial values of above-inversion total water mixing ratio

typical of cumulus cloud layers) towards large LWP than the high cloud fraction cases (high initial q_t^+) in the MIXED series. For lower initial q_t^+ , the distribution shows no maxima at the mean LWP value but an exponential decay towards extreme LWP with a noisy, long tail. This is additional evidence of transition from cumulus-into-stratus to cumulus as κ increases. There is also clear distinction between the low initial q_t^+ (low cloud fraction) cases in the MIXED and DECOUPLED series. Observational studies of low clouds usually show LWP distributions classified according to cloud fraction (e.g., Barker et al. 1996; Zhou et al. 2006; Wood and Hartmann 2006). The LWP distributions for high cloud fraction cases shown by these studies are similar to those found in the high cloud fraction cases in the MIXED and DECOUPLED series. The distinction between the high cloud fraction cases in the MIXED and DECOUPLED series in terms of the tail,

however, is not found in such averaged distributions over mixed cases. The low cloud fraction cases in the observational studies are usually associated with LWP distributions with an exponential decay, similar to those found in the low cloud fraction cases in the DECOUPLED series. The study of Barker et al. (1996) made a further distinction between broken stratocumulus and trade cumulus (by eye). In the cases classified by Barker et al. (1996) as broken stratocumulus we found one (scene B2) with similar LWP distributions to the low cloud fraction cases in our MIXED series.

4 Dependence of the results on model resolution and microphysics parameterization

The qualitative aspects of the transitional behavior from a cumulus-into-stratus regime to a shallow cumulus regime with increasing κ in the DECOUPLED series might be model independent, but the quantitative aspects might depend on LES characteristics such as resolution and microphysics parameterization. To address this issue, we repeat the DECOUPLED series (1) by doubling the horizontal resolution used so far (i.e., reducing the grid size from 100 to 50 m) and (2) by turning on the microphysics parameterization option for warm rain (see Seifert and Beheng 2001, 2006. The cloud droplet number concentration is set at 70 cm^{-3} following the setup of Lock (2009).) to allow for precipitation. The dependence of cloud fraction and liquid water path on increasing κ in these two new series of experiments are shown along side the original series in Fig. 8. The marked transition from high-cloud-cover cumulus-into-stratus regime to low-cloud-cover shallow cumulus regime is maintained in both the high resolution series and the precipitation-allowing series. The quantitative details of the transition change, however.

The higher resolution DECOUPLED case with initial $q_t^+ = 4.0 \text{ g/kg}$ now maintains a cumulus-into-stratus regime with cloud cover larger than 60% (compared to $\sim 35\%$ in the lower resolution experiment with the same initial q_t^+). This indicates the transition requires slightly larger κ in the higher resolution series. A comparison between the higher and lower resolution series suggests that with higher resolution, higher cloud cover can be maintained with smaller LWP. We could argue, therefore, that a stronger positive feedback between turbulence and radiation helps maintaining higher cloud cover in the higher resolution DECOUPLED series, which is consistent with Stevens et al. (2001). This stronger feedback is also consistent with transition happening at larger κ in the higher resolution series since the same level of turbulence mixing can be maintained with stronger entrainment drying (larger κ).

Fig. 8 The scatterplots for κ versus cloud fraction (on the left) and κ versus liquid water path (on the right). The values are all averaged over the last 3 h of the simulations. The triangles are cases in the original DECOUPLED series; the crosses are cases in the high resolution DECOUPLED series; and the squares represent cases in the DECOUPLED series with precipitation processes. The symbols are color coded according to the initial values of above-inversion total water mixing ratio

In the precipitation-allowing DECOUPLED series, the values of LWP for cases with small κ are much smaller than in the original DECOUPLED series without precipitation. The cloud cover is also reduced by precipitation but less dramatically than LWP. Precipitation changes the values of κ in this series, which is in general larger than in the series without precipitation. This is consistent with the results of Lock (2009). The response of cloud cover and LWP to changes in κ , however, is consistent with our previous conclusions.

5 Conclusions and discussion

5.1 Conclusions

Our LES results were able to reproduce the four scenarios in Table 1, thus confirming the motivating hypothesis of this study. Only in decoupled stratocumulus-topped marine boundary layers (SMBLs) with large κ (scenario (A)) did we find very high probability of transition to trade-wind cumulus topped marine boundary layers (TCMBLs). The essential difference between SMBLs in scenarios (A) and (B) is the larger surface evaporation in (A) (132 W m^{-2} compared to 96 W m^{-2}), which gives rise to a much deeper cloud layer sustained by cumulus updrafts, rather than by the stratocumulus and their radiative effects, as pointed out by WB97. Larger κ leads to a transition to TCMBL in (A) while it only leads to “deterioration” towards a surface heat flux driven MBL in (B). The difference between SMBLs in scenarios (A) and (C) is the positive feedback for stratocumulus destruction enabled by larger κ in the former, i.e., the much more suitable condition across the inversion for the SMBL-to-TCMBL transition. This result confirms our hypothesis that the combined influence of

thermodynamic conditions at the surface and right above the inversion on the SMBL-to-TCMBL transition is more effective than any one considered individually.

The evaporatively enhanced entrainment proposed by Siems et al. (1990) is not replicated in our results for either the DECOUPLED or the MIXED series. For the MIXED series, we obtain an increase in cloud patchiness as κ increases, which may be associated with the locally energized entraining eddies due to evaporatively enhanced entrainment. Our interpretation is that the effect of evaporatively enhanced entrainment in our simulations is shadowed by that of entrainment drying, which reduces the radiative cooling and hence entrainment. The evaporatively enhanced entrainment intensity could affect the efficiency of the positive feedback responsible for the transition to TCMBLs for SMBLs in scenario (A).

There are several recent studies on buoyancy reversal and CTEI similar to ours, e.g., Lock (2009) and Yamaguchi and Randall (2008). Lock (2009) studied the sensitivity of cloud fraction and LWP to κ with a LES setup similar to our DECOUPLED series while Yamaguchi and Randall (2008) studied the impact of buoyancy reversal with a LES setup similar to our MIXED series. Lock (2009) emphasized the destruction of cloud through cloud-top entrainment and suggested κ as an efficient control on cloud fraction in the decoupled MBL. He further pointed out that for small κ (< 0.2) uniform stratiform cloud cover ($> 80\%$) can be maintained while for large κ (> 0.5) only cumulus clouds ($\sim 20\%$) exists. Yamaguchi and Randall (2008) emphasized for well-mixed MBLs, the role of cloud building processes (e.g., radiative cooling, surface fluxes) in masking cloud destruction by entrainment (controlled by κ). Our results are consistent with both of these studies. Furthermore, we show that it is important to consider the internal structure of the MBL, i.e., well-mixed or decoupled, when estimating the response of low clouds to changes in the stability of the inversion. We argue from these results that it is important that we consider the large-scale control on the low clouds from both the surface and across the inversion when parameterizing the associated subgrid scale processes in large scale models.

5.2 Implications for MBL parameterizations to be used in GCMs

According to our results with the LES, parameters such as domain averaged κ , k and D , which were originally designed to measure the probability of buoyancy reversal in large-scale models, are good indicators of the “dryness” of the inversion. Actually, those parameters could be viewed as counterparts of the Bowen ratio at the surface. The larger the surface evaporation (smaller Bowen ratio), the drier the air above the inversion (large κ), the more

suitable it is for cumulus updrafts to develop in favor of stratocumulus.

Other parameters measuring the degree of decoupling in the SMBL (α_q , α_θ and r), describe the response of the SMBL to changes in the surface and free atmosphere conditions, but do not suffice to prognostically determine the transition to TCMBL. Our results show that for a decoupled SMBL, when the buoyancy reversal parameters (κ , k and D) are large enough, the transition to TCMBL is guaranteed. This gives us the possibility of parameterizing the transition to TCMBL for SMBLs in scenario (A) in Table 1. For SMBLs in scenario (B), a transition to TCMBL did not occur within the range of κ seen in our study. Large κ in this case may lead to decoupling, which is not caused by the development of surface-driven cumulus updrafts but rather by dissipation of cloudiness from above. This type of decoupling would not lead to a transition to TCMBL unless the surface conditions change too. As a first step to implement the above parameters in the mixed-layer model based boundary layer scheme in a GCM these parameters could be used as a “switch” to determine the boundary layer structure in the design of a boundary layer parameterization. When both conditions are satisfied (i.e., decoupled SMBL with large κ), the boundary layer starts venting cloudy air to the free atmosphere. This process tries to mimic the transport of moisture and heat by trade cumulus plumes in TCMBLs. The details of the algorithm could be further tested to assure the continuity between the two types of boundary layer. This algorithm is not meant to replace cumulus parameterizations, but to remedy the deficiencies of existing cumulus and boundary layer parameterizations in dealing with the transition.

The problematic scenario in terms of formulation is SMBLs in (C) in Table 1. As in the two-step conceptual model for the transition discussed by WB97, entrainment caused by impinging cumulus updrafts alone might be able to break the balance and destroy the overlying stratus as the SMBL moves over warmer water. It has also been shown by Stevens et al. (2001, see their Figure 9 and related discussions), however, that radiation-turbulence feedback in the stratocumulus layer very likely plays a critical role in the evolution of decoupled SMBLs. Thus, in scenario (C) we must take into account interacting turbulent, radiative and microphysical processes in both the shallow cumuli and overlying stratus. This remains a tough problem whether we explicitly parameterize the transition or incorporate these processes through coherent parameterization assumptions of the boundary layer. The importance of this problem has been demonstrated by recent observational studies (e.g., Wood and Bretherton 2004; Albrecht et al. 1995), according to which decoupled SMBLs might be very spread over subtropical oceans. For future improvement of boundary layer parameterization in climate

models, it might be beneficial to evaluate the importance of errors induced by misrepresenting SMBLs in scenario (C).

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