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Influence of seasonal variation in cloud condensation nuclei, drizzle, and solar radiation, on marine stratocumulus optical depth

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ABSTRACT

A modeling study was performed to investigate the factors that control the seasonal cycle in the optical depth of marine stratocumulus cloud at a Southern Hemisphere site. Cloud optical depth is primarily influenced by cloud depth and cloud droplet concentration, the latter being controlled by more or less known fluctuations in cloud condensation nucleus (CCN) concentration. The study primarily focussed on the factors influencing cloud depth. The results show that there are two important factors controlling cloud depth, namely solar radiation and drizzle. Seasonal variations in insolation act to enhance cloud depth in winter compared to summer, while drizzle acts in the opposite direction. Drizzle intensity is calculated to be higher in winter when the concentration of CCN is low and cloud droplet effective radius is large. Subcloud evaporation of drizzle droplets reduces entrainment rate, while increased precipitation at the surface depletes cloud water. Thus, because seasonal cycles in cloud depth due to variations in insolation and drizzle are of opposite phase the result is a partial cancelation of their separate influences on cloud optical depth.

1. Introduction

It is recognized that clouds are poorly represented in numerical models of the atmosphere. However, not only are clouds one of the largest modulators of climate but feedbacks related to them currently cause perhaps the largest uncertainties in prediction of climate change. Marine stratocumulus clouds are responsible for a radiative forcing on the earth's climate system which is of opposite sign to the conventional greenhouse radiative forcing resulting from increased emission of carbon dioxide and associated increases in tropospheric water vapor (Ramanathan et al., 1989). Therefore, it is of great interest to quantify the radiative structure of these types of clouds.

One topic of study in the research community is the link between cloud condensation nuclei (CCN) and cloud albedo which was postulated many years ago (Twomey, 1977), in the context of anthropogenic injection of CCN into the atmosphere. The principle idea was that increased emission of CCN caused by worldwide pollution would increase cloud albedo, partly offsetting the conventional greenhouse effect. However, Charlson et al. (1987) in a related hypothesis suggested the existence of a natural regulation of cloud albedo caused by the emission of dimethylsulfide (DMS) from the ocean, which in the atmosphere is a major precursor to CCN.

We can treat the theory that perturbations in CCN will lead to a perturbation in cloud albedo through a direct effect on cloud droplet concentration as highly likely as there is substantial evidence to support it (Radke et al., 1989). Whether such perturbations are due to anthropogenic or natural causes is immaterial although most efforts to prove the theory have addressed the anthropogenic aspects. Variability in cloud albedo due to natural effects such as DMS-emission are much harder to prove as signals are weaker, and many intermediate links between DMS, CCN-precursors and CCN need to be confirmed.

Boers et al. (1994) have shown for a Southern

Hemispheric site a coherence in seasonal variation between satellite derived cloud optical depth and surface based observations of CCN concentration. They indicated that, although the seasonal variation in optical depth as observed from satellite measurements is widespread, coherence between these cycles and cycles in DMS is by no means assured since many factors contribute to cloud optical depth, of which a natural variability in CCN due to natural perturbations in DMS is only one.

They used an adiabatic model of cloud droplet growth on CCN to predict the droplet number concentration from which a seasonal variation in cloud optical depth could be obtained under the assumption that cloud depth was constant. However, they indicated that (a) the result was quite sensitive to this assumption, and (b) there is probable cause to believe that the seasonal variation in cloud depth would contribute to a seasonal cycle in cloud optical depth with opposite phase to the seasonal cycle in optical depth due to CCN alone. This last conclusion was based on a model study in which solar radiation was varied seasonally.

The purpose of this paper is to explore the seasonal variability in cloud depth from a modelling perspective. It is stimulated by our desire to understand the principle mechanisms contributing to observed cycles in cloud optical depth in addition to the obvious direct link between CCN and cloud droplet concentration.

The next section contains a brief description of the model and approach used. Then, the results of the simulations are presented from which it is evident that the important regulators of cloud depth are solar radiation and drizzle. The simulation is performed using inputs for sun angle and cloud droplet concentration that are representative for the Southern Hemisphere site considered by Boers et al. (1994) (Cape Grim, Tasmania, 40°41′S, 144°41′E).

2. Boundary layer model

2.1. Background

In the last 25 years, a considerable effort has gone in modeling the atmospheric boundary layer under clear and cloudy conditions. The work of Lilly (1968) was responsible for a great number of papers on parameterizing marine boundary layer processes in order to predict the entrainment rate at cloud top responsible for regulating boundary layer depth and cloud depth.

His model and others start from the principle that the marine boundary layer over the temperate oceans is well-mixed with conservative variables such as total water mixing ratio (q_T) and equivalent potential temperature (θ_e) constant with height. The fundamental implication of such simple profiles is that the buoyancy fluxes driving convective overturning and entrainment are constrained to be linear with height so that entrainment can be predicted using only one assumption about the buoyancy production of turbulence inside the layer. The simplifying assumption of a well-mixed boundary layer is subject to some debate, as there have been studies in support of this assumption (Brost et al., 1982; Stage and Businger, 1981), as well as studies which do not support it (Nicholls, 1984). Despite this uncertainty, the use of the assumption of a well mixed state is very useful in describing the fundamental evolution of a cloudy marine boundary layer.

The assumption of a well-mixed boundary layer clearly does not hold for the trade wind cumulus regime leading to modeling attempts which separated the cloud layer from the subcloud layer (Albrecht et al., 1979). Using this scheme the buoyancy fluxes in the cloud layer need to be predicted with a mass flux model of updrafts and downdrafts. Furthermore, cloud fraction can be predicted.

When for specific applications the details of the flux profiles need to be known it is necessary to invoke second-order and even third-order closure schemes of the turbulence quantities in the boundary layer (Deardorff, 1980; Andre et al., 1978).

The price paid for the increased complexity in model formulations is a substantial increase in the number of uncertain parameterizations necessary to make models work in a realistic way. In order to study the influence of cloud depth on cloud optical depth such complexity is probably not necessary, hence the decision to use the assumption of a well mixed state as advanced by Lilly. The model closure to be used in this study is a derivative of Lilly's (1968) work by Turton and Nicholls (1987).

2.2. Model assumptions

The principle assumption in models of the Lilly type is that the conservative variables (θ_e, q_T) are invariant with height with an infinitesimal jump in these variables at the top of the layer where the air is in contact with the overlying atmosphere. Under stationary conditions the top of the layer is maintained at an altitude as a result of the competing effects of subsidence which tends to lower it, and the entrainment rate which envelops the overlying air into the boundary layer. In a model of this type the entrainment rate is unknown; its knowledge would specify the thermodynamic fluxes across the interface, which together with known surface fluxes constrains the fluxes throughout the depth of the layer.

Since there is general agreement that the intensity of turbulence controls the entrainment rate, it is usually parameterized as a fraction of the (buoyant) production of turbulence in the interior of the layer. Exactly how much of the turbulence kinetic energy (TKE) is converted into energy of entrainment is not known, but has given rise to a large volume of papers debating the merits of various assumptions, none of which have been thoroughly tested against observations (Randall, 1980; Randall, 1984; Stage and Businger, 1981). However, most paramerizations do constrain the depth of the layer to heights commonly found over the temperate oceans (0.8–1.5 km).

The Turton and Nicholls' (1987) model uses a Richardson number dependence of the entrainment rate with an empirical adjustment that constrains the entrainment under conditions when the buoyancy jump across the inversion top is small:

$$w_e/w_* = a_1 \mathrm{Ri}^{-1}(1 + a_2(1 - \Delta_m/\Delta\theta_v)), \tag{1}$$

where w_e the entrainment rate, w_* the convective velocity scale, a_1, a_2 are constants, $\Delta\theta_v$ the virtual potential temperature jump across the inversion, and Δ_m the adjustment term (Nicholls and Turton, 1986). The Richardson number Ri is defined as:

$$\mathrm{Ri} = g z_b \, \Delta\theta_v/(\theta_v w_*^2), \tag{2}$$

where g the gravitational acceleration, and z_b the boundary layer depth.

Many similar models (Randall, 1980; Stage and Businger, 1981) lack the adjustment term Δ_m. The absence of this term can give rise to unrealistic growth rates.

It should be noted that the Turton and Nicholls' model additionally considers a decoupling stage for which it is probably better known than the entrainment formulation. In the decoupling stage, the upper part of the boundary layer (which includes the cloud) becomes separated from the surface layer by a small buoyancy jump. The evolution of the buoyancy jump is governed by the intensity of turbulence in the cloud layer. Simulations performed with their model and including the decoupling stage revealed several unrealistic features, one of which was that the buoyancy jump separating the surface boundary layer from the cloud layer restricted the vertical transport of moisture from the ocean surface too much. As a result clouds often developed in the surface boundary layer, which is not supported by observations.

Most of the problems with the decoupling stage can be reduced to the fact that the process of decoupling is inadequately described by the development of a buoyancy jump only. On occasion, the conservative variables develop gradients with height (Boers and Betts, 1988) when the turbulence intensity is insufficient to keep the layer well-mixed, but still sufficient to inhibit the formation of a decoupling buoyancy jump.

In this study we therefore only retain the entrainment formulation (1). This limits the applicability of the results to conditions where a well-defined decoupling stage such as described by Nicholls (1984) is absent.

2.3. *Radiative transfer*

Over the oceans, most of the generation of turbulence kinetic energy is caused by cloud-top cooling in the infrared region of the spectrum. Buoyancy fluxes associated with cloud-top cooling are as large as the fluxes associated with surface heating of a continental boundary layer on a sunny morning (60–100 W m^{-2}) (see for example Randall, 1980).

On the one hand, solar radiation restricts the rate of cooling because of absorption of radiation by cloud droplets and enhanced water vapor absorption of radiation caused by multiple scattering inside the cloud. On the other hand, unlike the process of infrared cooling, solar warming is not confined to the upper 100 m of the cloud, but spread out over a larger depth, which results in a local source of turbulence. However, in all cases the decrease in cloud-top cooling outstrips the local generation of turbulence as a dominant influence on entrainment with the result that during the daytime the depth of the boundary layer is reduced. The reduction of boundary layer depth, combined with boundary layer heating, reduces overall cloud depth.

Most studies have examined the influence of solar radiation on cloud development in the context of diurnal variations in boundary layer structure (Bougeault, 1981; Nicholls, 1984; Blascovich et al., 1991; Albrecht et al., 1990) although an occasional study has hinted that seasonal variations in insolation could have significant influence on cloud depth as well (Hanson and Gruber, 1982).

The importance of solar radiation in controlling the generation of turbulence kinetic energy merits the inclusion of a detailed radiative transfer model in the region of the solar spectrum into the model that describes the overall boundary layer evolution. The approach adopted here is that used by Slingo and Schrecker (1982). It divides the solar region into 24 spectral bands where either water vapor, or ozone are the dominant absorbers. The scattering properties (g, the asymmetry parameter, σ the extinction, ω, the single scatter albedo) of the cloud droplet distributions in each spectral band were parametrized in terms of the effective radius of the cloud droplets. The radiative cooling and heating rates at the longwave end of the spectrum are parameterized with a computationally fast algorithm based on the work of Harshvardhan et al. (1987).

All simulations were performed by using a diurnal cycle in the sun angle (at the latitude, longitude of Cape Grim, 40°41′S, 144°41′E, and indicated month). A simulation using a diurnal cycle in sun angle is preferred over the use of daily averaged sun angle which is often used in climate studies (see f.e. Betts and Ridgway, 1988). Test simulations showed that the two are not interchangeable. The reason is that the diurnal cycle in solar absorption causes a diurnal variation in the depth of the cloud, which in turn affects solar absorption, and thus boundary layer development. When a constant sun angle is used in the simulation the diurnal variation in cloud depth is absent. Thus, integrated over time the total solar absorption is different as well so that differences in averaged optical depth of up to 15% occur between the two types of simulations.

2.4. *Drizzle*

The recognition that drizzle can significantly affect the overall development of the cloudy boundary layer is a recent advancement in the study of boundary layer clouds. Drizzle rates are controlled by the size of cloud droplets near cloud top, with the conversion process the highest for large droplets. Albrecht (1989) suggested that any perturbation in CCN would affect the drizzle process and thus control cloud amount. Further work by Wang (1993) suggested an influence of drizzle on cloud depth with drizzle generally lowering cloud depth when compared to clouds that do not precipitate.

The present choice of model formulation does not merit the inclusion of a mechanism to describe the drizzle autoconversion process (Chen and Cotton, 1987; Baker, 1993) since the main interest is in the effect of a drizzle parameterization on boundary layer development, rather than judging which parameterization is the best. The evaluation of any drizzle parameterization is hazardous at best, since such an exercise would imply accurate precipitation measurements, and observations of the (small) tail of large (precipitable) water droplets of the cloud droplet spectrum, both of which are rarely available.

It is not unrealistic to assume that drizzle generation starts at cloud top, and that coalescence collects smaller droplets on larger droplets as they falls towards the bottom of the cloud. Based on few observations Turton and Nicholls (1987) assumed that the drizzle flux was constant throughout the cloud, while the evaporation rate was constant below the cloud. Thus, if the evaporation rate below the cloud, and the drizzle flux at cloud top are specified, the water budget, and hence the buoyancy fluxes can be calculated. Their formulation for the drizzle flux at cloud top is retained in this study:

$$F_{\mathrm{dr}} = -3 \times 10^{-5} \alpha \, \mathrm{LWP}^{1/2} \qquad (3)$$

where LWP the liquid water path. The factor α ($0 \leqslant \alpha \leqslant 1$) can be viewed as an efficiency factor which for the purpose of this study was set to be linearly dependent on the mean volume radius at cloud top: $\alpha = r_v/20$ ($r_v \leqslant 20\,\mu$m), $\alpha = 1$

($r_v > 20\ \mu m$), and r_v the mean volume radius. This is a somewhat weaker efficiency than proposed by Turton and Nicholls (1987) based on the fact that the loss of large particles from cloud top must reduce the efficiency with which cloud droplets can be converted to drizzle droplets. Below cloud base a fixed rate of decrease in flux was specified as 0.86×10^{-5} m s^{-1} km^{-1}, as in Turton and Nicholls (1987).

The influence of a drizzle process on the water budget and turbulence kinetic energy budget of the boundary layer are profound (Nicholls, 1984). Drizzle rates are of the same order of magnitude as the evaporation rate at the surface, as any simple calculation will attest. Consequently, it will be hard for a cloud to maintain its liquid water since the presence of large particles will evoke an instantaneous response on the vertical water flux. Furthermore, the evaporation of drizzle below the cloud suppresses turbulence and acts to stabilize the subcloud layer. A stabilization of the subcloud layer is not taken into account in the model used here because it assumes well-mixed conditions throughout. However, the diurnal cycles in turbulence energy production as calculated by the model clearly indicate a reduction in turbulence kinetic energy generation during the daytime hours. The suppression of turbulence comes about because the apparent water vapor flux which contributes to the buoyancy flux is reduced by the downward flux of water in the form of drizzle droplets. Especially below cloud, this leads to a region of strong negative buoyancy flux, which competes for turbulence kinetic energy with the region of negative buoyancy flux at the inversion top to maintain the well-mixed state. As a result entrainment decreases and the boundary layer top sinks.

This type of model for the drizzle flux was recently used by Pincus and Baker (1994) to study the albedo suspectibility of clouds to the drizzle process. Their results clearly show the sensitivity of cloud depth (and hence optical depth) to changing CCN which through its effect on droplet size and concentration affects the drizzle process.

3. Model calculations

3.1. *Strategy*

In this study the seasonal variability of marine stratocumulus clouds is examined at the temperate latitudes where they are most common. Although most efforts have been directed at typical cloudy conditions off the coast of California, the present focus is on the Southern Ocean south of Australia. Stratiform clouds are quite common there as well; furthermore, there is the additional benefit of modeling a region just upwind of the Cape Grim Air Pollution Baseline Station (CGBAPS) for which more than a decade of CCN concentration data are available (Gras, 1989).

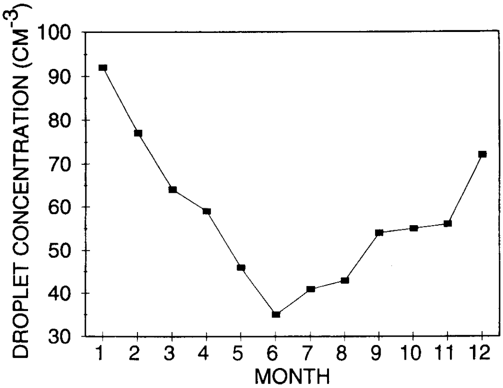

Fig. 1. Modeled cloud droplet concentration used as input in the mixed-layer model computations.

A midlatitude summertime thermodynamic profile with 30% relative humidity was used as an initial condition. Sea surface temperature and the surface divergence rate were fixed at 14°C, 6.5×10^{-6} s^{-1}, respectively, while the monthly values of cloud droplet number concentration were calculated by a CCN, cloud droplet conversion model (Ayers and Larson, 1990) based on CCN, data as observed at Cape Grim (Gras, 1989). The input values of cloud droplet concentration (Fig. 1) used in this simulation are thus exactly those of Boers et al. (1994) (who used an updraft speed of 20 cm s^{-1}), ranging from 35 cm^{-3} in winter to 92 cm^{-3} in summer. Although there are undoubtedly seasonal variations in both the divergence rate, and atmospheric stability, these parameters were left unchanged as the sensitivity of model predictions to those parameters have been treated elsewhere (Randall and Suarez, 1984; Driedonks, 1982). Wind speed was fixed at 10 m s^{-1}.

Using these initial conditions the model was run near the middle of each month until stationarity

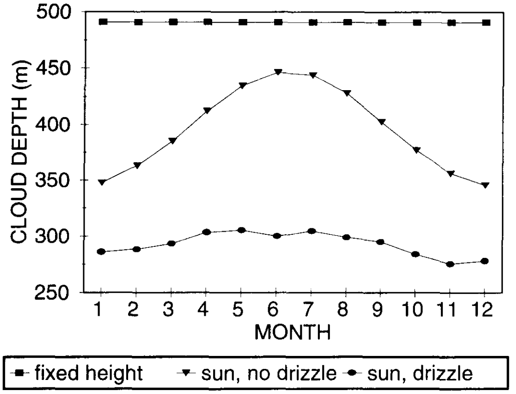

Fig. 2. Seasonal cycles in cloud depth.

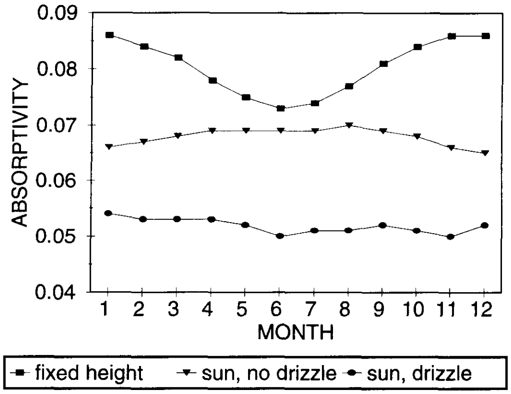

Fig. 4. Seasonal cycles in cloud absorptivity.

was reached (approximately 8 days). At the end of the model runs various model parameters were recorded for intercomparison between different model runs.

The strategy employed has been to initially perform a baseline simulation with fixed height, and no drizzle (*Simulation* 1). Solar radiation was allowed to vary seasonally, but only to compute cloud absorptivity and reflectivity, not to compute the internal response of the boundary layer. In *Simulation* 2, boundary layer entrainment rate and conservative variables were allowed to vary in response to the solar input, while drizzle and solar radiation were combined in one model run to study their joint influence in *Simulation* 3. In all instances the external conditions remained the same.

The results of *Simulation* 1, 2 and 3 are shown in Fig. 2 (cloud optical depth), Fig. 3 (cloud depth), Fig. 4 (cloud absorptivity), Fig. 5 (cloud reflected flux), and Fig. 6 (flux albedo). *Simulation* 1 describes the classical variability of cloud optical depth as as result in variation of cloud droplet number concentration by conforming to most studies by simply extrapolating the variation in cloud droplet number concentration to cloud optical depth leaving cloud depth constant. As will be shown in *Simulation* 2, this is an oversimplification with serious implications because solar radiation controls cloud depth, hence cloud optical depth. Boers and Mitchell (1994) hinted at this feedback on cloud optical depth, although they did not perform a full interactive boundary layer model calculation.

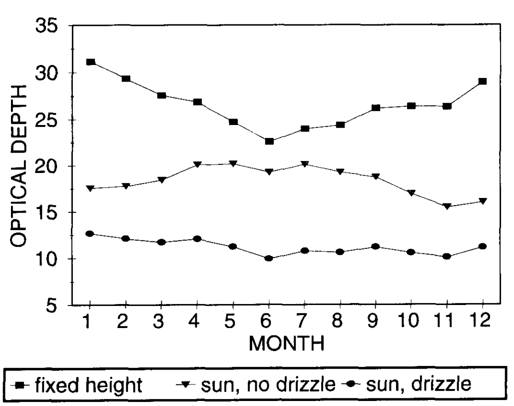

Fig. 3. Seasonal cycles in cloud optical depth.

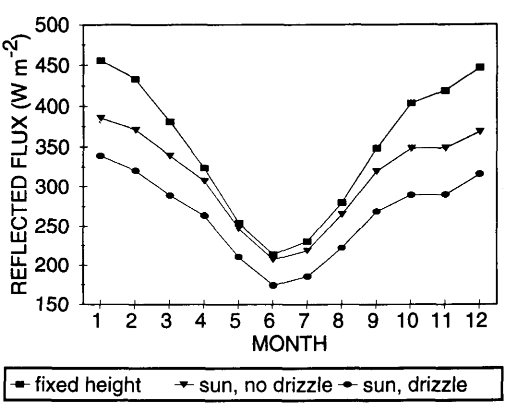

Fig. 5. Seasonal cycles in cloud reflected flux.

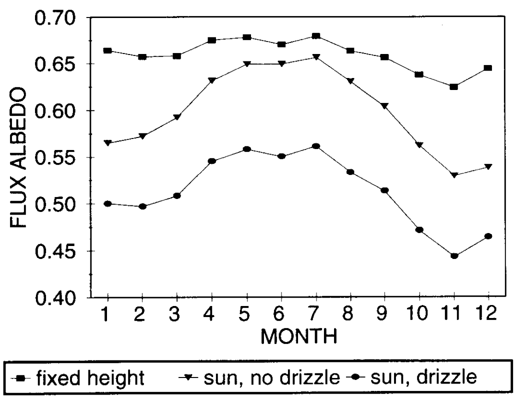

Fig. 6. Seasonal cycles in cloud top albedo.

3.2. *Fixed height, no drizzle, Simulation 1*

Simulation 1 does not allow for seasonal variation in cloud depth (Fig. 2). However, it does allow for changes in optical depth (Fig. 3), ranging from 30 in January to 23 in June, as its variability is entirely controlled by seasonal changes in cloud droplet number concentration.

The absorptivity of the cloud for solar radiation changes from 8.5% in the summer months to 7.2% in the winter, solely the result of the changing angle of insolation. Although the reflected flux at cloud top varies strongly between 450 W m^{-2} in summer to 220 W m^{-2} in winter, the flux albedo is the highest in May and July (67%), due to the importance of the solar zenith angle in determining cloud albedo.

3.3. *Changing height, Simulation 2*

When solar heating is allowed to modify the turbulence energy budget and the thermodynamic budget of the boundary layer, it produces a dramatic reduction in cloud depth. Cloud depth is largest in June–July (450 m, austral winter) and smallest in January (350 m, austral summer) (Fig. 2).

Cloud optical depth is now a function of cloud depth as well as cloud droplet number concentration. The results indicate that for this simulation the change in cloud depth dominates the seasonal optical depth variability, because the optical depth has a maximum in winter (20) and a minimum in summer (16). Cloud absorptivity is almost constant for this simulation. Apparently, the cloud strives towards an absorption equilibrium, which is maintained despite large changes in insolation and cloud depth. The reason for this absorption equilibrium is that the entrainment rate is a strong function of cloud absorptivity. In any season an equilibrium height is maintained by a balance between subsidence rate and entrainment rate. The entrainment of warm air into the boundary layer compensates for the loss of heat due to cloud top cooling. In the winter season, with low levels of insolation, a thicker cloud is necessary to absorb the solar radiation necessary to balance the subsidence rate, while in the summer season, with high levels of insolation, the opposite occurs. Cloud albedo now peaks in mid winter at 65% and has a minimum in November (44%).

3.4. *Drizzle, and solar radiation, Simulation 3*

If, in addition to solar radiation, a parameterization of the drizzle process is allowed to modify the turbulence kinetic energy budget, cloud depth and optical depth are further reduced. Since the droplets are the largest in winter, the drizzle process is the most intense with the result that the reduction in wintertime cloud depth is the largest (150 m). It is reduced by only 60 m in the summertime. Clearly, the effect of drizzle is to counteract the seasonal variations in cloud (optical) depth induced by solar radiation alone. The amount by which cloud depth is reduced is undoubtedly dependent on the manner in which drizzle is parameterized. However, in all cases the overall effect of a reduced cloud depth is the same.

Since the cloud depth variability due to solar radiation and that due to drizzle are 6 months out of phase it comes as no surprise that the overall cloud depth remains about the same apart from some small month to month variations. Once again it is clear that the cloud seeks to maintain an absorption equilibrium, albeit at a lower value than for *Simulation* 1. The cloud albedo is also reduced varying from 44% in November, to 56% in July.

Fig. 7 shows the modeled liquid water flux due to precipitation (negative downward). This flux goes through a strong seasonal cycle with the least precipitation in summer, and the most in winter. The computations are in qualitative but not quantitative agreement with the 15 year precipitation record at Cape Grim.

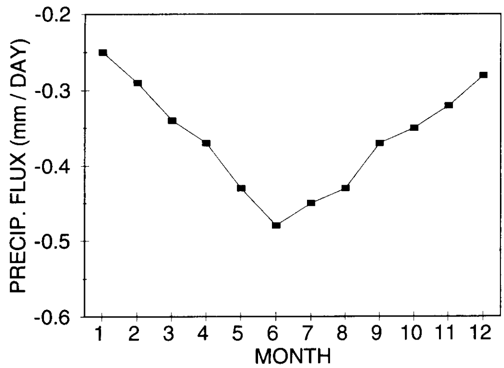

Fig. 7. Seasonal cycles in liquid water flux due to precipitation.

4. Discussion and conclusion

The large reduction of cloud optical depth from Simulation 1 to Simulation 3 would seem to be a significant result. Whilst the divergence rate was fixed for all simulations, solar radiation and drizzle to a very large extent appear to be responsible for maintaining the cloud optical depth at values of 10 to 15 which often are reported in the literature. In light of this result the assumption of Boers et al. (1994) of constant cloud depth in their study linking the seasonal variability of CCN to cloud optical depth is not unrealistic. However, this constant depth seems to be the result of two compensating mechanisms, namely drizzle and solar radiation.

In assessing the strength of both drizzle and solar radiation in reducing the yearly cloud albedo it was found that if cloud depth is allowed to remain constant year around, the calculated average yearly albedo of the cloud is 0.66; if cloud depth is allowed to change in response to solar heating, this albedo is reduced to 0.60, and if drizzle is included it is further reduced to 0.52. Thus the inclusion of drizzle more than doubles the yearly albedo reduction. Note also that this result is obtained with the assumption that the boundary layer remains well-mixed, even if drizzle rates are pronounced. This is probably the weakest assumption of the type of model used here. Furthermore, the depletion of liquid water due to precipitation will remove a portion of small cloud particles that are essential in determining the reflective capability of the cloud.

For Simulation 3 the cloud optical depth was found to be lowest in the wintertime, while the cloud albedo is the highest. Although this result is latitude dependent due to the variability of the sun angle it highlights the fundamental difference between these two parameters, one of which (cloud optical depth) is strictly a function of the local cloud vertical dimension and microphysical structure, while the other (albedo) is an additional function of sun angle. Since albedo is a non-linear function of optical depth, it follows that seasonal perturbations in cloud optical depth affect the yearly averaged cloud albedo in a non-linear way, too.

This study indicates that seasonal variability in cloud optical depth directly due to variations in CCN–DMS concentration can be easily masked if the two compensating mechanisms are slightly out of phase, or if one dominates the other. Such phase shifts can occur if for example the large scale subsidence rate which was held constant for the purpose of this study undergoes significant seasonal variations. Furthermore, there are seasonal changes in both sea surface temperature, and atmospheric stability which may play a role in modifying the seasonal variation in cloud optical depth.

Some caution is needed to generalize the results of this study to other areas of the globe. The model results are very dependent on the amount of solar radiation affecting the cloud. Since the amount of solar radiation is dependent on latitude, it follows that optical depth variations due to changing sun angle are dependent on latitude as well.

Most importantly, this study indicates that great care needs to be taken in estimating the effect of CCN on cloud albedo. Given the results presented here, the dynamic adjustment of cloud depth to changing external conditions is an important element which up to now has received little attention.

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