Boundary Layer Clouds in a Global Atmospheric Model: Simple Cloud Cover Parameterizations

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ABSTRACT

Subtropical boundary layer clouds have a fundamental role on the radiative budget of the atmosphere and on the modulation of the tropical circulations. The development of realistic parameterizations of these clouds in global atmospheric models is a major challenge. Unfortunately, this has been a difficult problem to solve in an acceptable way. In this paper, new and simple parameterization schemes for subtropical clouds are implemented in the U.S. Navy Operational Global Atmospheric Prediction System’s (NOGAPS) forecast model. The parameterizations are partially based on large eddy simulation (LES) results and provide a substantial improvement when compared with observations and with the previous scheme. The global distribution of boundary layer clouds and of surface shortwave radiation is more realistic with this new scheme, particularly over the stratocumulus regions. The transition from stratocumulus to cumulus is well captured, the seasonal and diurnal cycles of stratocumulus are realistic, and the Arctic boundary layer temperature is improved with the new parameterizations.

1. Introduction

The occurrence of stratocumulus and shallow cumulus is a prominent feature of the climate system (e.g., Klein and Hartmann 1993; Siebesma 1998). The large values of stratocumulus cloud cover are usually associated with cold sea surface temperatures and atmospheric subsidence. Because of its high albedo, stratocumulus has a significant cooling effect on the ocean. It has been recently suggested that stratocumulus play a significant role in determining the tropical and subtropical atmospheric circulation (Philander et al. 1996; Ma et al. 1996; Larson et al. 1999). On the other hand, shallow convection has a significant influence on the large-scale circulation in both the Tropics and the mid-latitudes (Siebesma 1998). Subtropical boundary layer (BL) clouds have, in general, a large impact on the BL turbulent structure.

A realistic representation of BL clouds in global atmospheric models has been a major issue in climate modeling and parameterization for a long time (e.g., Randall et al. 1985). There are today three major types of cloud parameterizations for climate and numerical weather prediction (NWP) models: diagnostic cloud schemes (e.g., Slingo 1987), where the cloud fraction is diagnosed as a function of relative humidity and some other parameters; statistical schemes (e.g., Mellor 1977; Sommeria and Deardorff 1977; Smith 1990), where the cloud fraction and liquid/ice water are diagnosed based on assumed probability distributions for the subgrid variability of the thermodynamical properties; and prognostic cloud schemes (e.g., Sundqvist 1988; Tiedtke 1993), where the liquid/ice water and the cloud fraction can be determined prognostically.

In spite of great advances in the development of more sophisticated cloud schemes in the last few years (e.g., Tiedtke 1993; Fowler et al. 1996; Del Genio et al. 1996), subtropical BL clouds are often not realistically represented in global models, and stratocumulus clouds, in particular, are often strongly underestimated (e.g., Duynerke and Teixeira 2001). As a consequence the shortwave radiation flux into the ocean surface is overestimated in these areas. In coupled ocean–atmosphere models, used for seasonal and climate prediction this could lead to large positive sea surface temperature (SST) biases (e.g., Stockdale et al. 1998).

In earlier studies, the Navy Operational Global Atmospheric Prediction System (NOGAPS) was coupled to the Modular Ocean Model (MOM) (Li and Hogan 1999). Coupled studies with the NOGAPS–MOM system indicated that stratocumulus is underestimated, leading to an excessive solar irradiance and consequently too warm SSTs.

Recently, a strategy involving the use of cloud re-
solving models (CRMs) or large eddy simulation (LES) models to evaluate and develop cloud parameterizations for large-scale models of the atmosphere has been pursued by the modeling community in general and, in particular, by the Global Energy and Water Cycle Experiment (GEWEX) Cloud System Study (GCSS; e.g., Bretherton et al. 1999) and the European Cloud Resolving Model (ECREM) project (e.g., Gregory and Teixeira 1998). In this paper, we propose two simple cloud cover prediction schemes for cumulus and stratocumulus. One is based on the steady-state version of the prognostic cloud fraction equation suggested by Tiedtke (1993). The other is based on a simplified version of the cloud scheme suggested by LES studies from Cuijpers and Bechtold (1995).

These two formulations are described in section 2, and are combined as a new cloud scheme. A validation of the capabilities of the new cloud scheme in producing a realistic global distribution of low clouds and of surface shortwave radiation is shown in section 3. Section 4 evaluates the performance of the new scheme in representing the transition from stratocumulus to cumulus and section 5 shows the improvements in the Arctic boundary layer temperature. The performance of the model in simulating the seasonal and diurnal cycles is presented in section 6 and vertical resolution issues are discussed in section 7. A summary is presented in section 8.

2. The model and the cloud scheme

a. The NOGAPS model

The NOGAPS forecast model (Hogan and Rosmond 1991) is a global spectral model in the horizontal with energy-conserving finite differences (hybrid-sigma coordinate) in the vertical. The dynamics formulation uses vorticity and divergence, virtual potential temperature, specific humidity, and terrain pressure as the dynamic variables. The model is centered in time with a semi-implicit treatment of gravity wave propagation. There is also wavenumber-dependent fourth-order diffusion of vorticity, divergence, and virtual potential temperature. The physics parameterization package includes: Richardson number–dependent boundary layer vertical mixing (Louis et al. 1982), surface flux parameterization (Louis 1979), gravity wave drag (Palmer et al. 1986), shallow cumulus mixing of moisture, temperature, and momentum (Tiedtke 1984), deep cumulus parameterization (Emanuel and Zivkovic-Rothman 1999), convective and stratiform cloud parameterization (Slingo 1987), and shortwave and longwave radiation (Harshvardhan et al. 1987).

In order to parameterize the interaction between clouds and longwave radiation, NOGAPS uses what is usually referred to as the broadband flux emissivity method (e.g., Stephens 1984). In it, the liquid water path is calculated based on an in-cloud liquid water content proportional to the saturation specific humidity, with a constant of proportionality that is 0.0079 for stable clouds and 0.056 for cumulus clouds. For the ice clouds the ice water content is estimated based on a temperature-dependent formulation from Heymsfield and Platt (1984). The mixed phase is taken into account assuming a linear mixture of ice and liquid from −40°C to 0°C. The shortwave optical thickness is calculated using the same liquid/ice water path as described for the longwave and a mean effective radius of 5 μm for stable clouds and 23 μm for the convective clouds.

The deep convection parameterization (Emanuel and Zivkovic-Rothman 1999) is based on the buoyancy-sorting concept and calculates the cloud-base mass flux in a prognostic way. A more realistic planetary boundary layer (PBL) structure was one of the major improvements, when compared with the previous relaxed Arakawa–Schubert scheme, provided by the new convection scheme.

b. A diagnostic cloud fraction scheme for cumulus

Prognostic cloud fraction schemes have been recently developed (Tiedtke 1993; Randel 1996; Wang 1996), and the one developed at the European Centre for Medium-Range Weather Forecasts (ECMWF) has been shown to realistically represent clouds in general (Tiedtke 1993; Jakob 1999; Morcrette et al. 2000) and the cloudy boundary layer in particular. However, it tends to underestimate stratocumulus and slightly overestimates cumulus (Jakob 1999; Teixeira 1999a; Duykerke and Teixeira 2001). It has been shown (Teixeira 2001) that one of the most fundamental aspects of its success is the direct coupling of the cloud scheme with the convection scheme.

Following one-column model studies (Teixeira 1999b), it was verified that, in cumulus situations, the cloud fraction prognostic equation is dominated by two terms:

\[
\frac{\partial a}{\partial t} = D(1 - a) - \frac{a^2}{l} K(q_s - q),
\]

where \(a\) is the cloud fraction, \(D\) is the detrainment rate, \(l\) is the liquid water, \(q\) is the specific humidity, \(q_s\) is the saturation-specific humidity, and \(K\) is an erosion coefficient. The first term on the rhs represents the production of clouds by detrainment from shallow convection and the second term represents the erosion of clouds by turbulent mixing with the environment.

In typical shallow convection situations the two terms on the rhs of the equation are of similar magnitude and the cloud fraction tendency is small compared to them. In this case, an equilibrium solution can be obtained from the steady-state version of Eq. (1):

\[
a^2 \frac{K}{l} q_s(1 - RH) - D(1 - a) = 0,
\]

where RH is the relative humidity.
Equation (2) is a quadratic equation in the unknown \( a \). Solving and neglecting the unphysical root, a diagnostic relation that gives cloud fraction as a function of \( \text{RH}, \ l, \ q_s, \) and \( D \) can be obtained (Teixeira 2001). However, a simplification can be made by using the in-cloud liquid water \( l_c \) instead of the liquid water. In this case, Eq. (2) simplifies to a first-order equation in the variable \( a \). Another reason to choose \( l_c \) is that a simple linear relation can be assumed between \( l_c \) and \( q_s \), which has been used in some models (e.g., ECMWF 1991) to predict the cloud water.

So, using \( l_c = lla \) and assuming \( l_c = \beta q_s \), we get

\[
a = \frac{D}{D + \frac{K}{\beta} (1 - \text{RH})}.
\]

This equation parameterizes, in a simple diagnostic way, the cloud fraction produced by the shallow convection scheme. Also, with this equation, the behavior of cloud fraction as a function of variables like the relative humidity, the detrainment, or the erosion coefficient, can be studied in a simplified manner. It is quite straightforward to see from Eq. (3) that \( \text{RH} = 1 \Rightarrow a = 1 \) and \( \text{RH} \sim 0 \Rightarrow a \sim 0 \) (as long as \( K/\beta \gg D \), which will always be the case for the present work).

A simple offline test illustrates the general behavior of the scheme. Equation (3) is used with the detrainment and the erosion coefficients specified as constant values: \( D = 5 \times 10^{-6} \text{ s}^{-1} \) and \( K = 5 \times 10^{-6} \text{ s}^{-1} \). These values will also be used on the NOGAPS implementation of the scheme. The detrainment value corresponds to the average value for shallow convection situations, obtained from an ECMWF simulation of the Atlantic Stratocumulus Transition Experiment (ASTEX) Lagrangian I (Teixeira 2000). The value for the erosion coefficient has been recently used in the ECMWF operational NWP model (e.g., Jakob et al. 1999). The inverse timescale \( 1/K \) provides information on the speed of the cloud’s dissipation. Using \( K = 5 \times 10^{-6} \text{ s}^{-1} \), \( \text{RH} = 0.8 \) and \( \beta = 0.01 \), corresponds to an \( e \)-folding time for the cumulus cloud dissipation of about 3 h.

Figure 1 shows values of cloud fraction as a function of \( \text{RH} \) for three realistic values of \( \beta = l_s/q_s \). It can be seen that the cloud fraction grows with larger values of \( \beta \). From analyzing Eq. (3) it can be seen that, in terms of cloud fraction, an increase in the parameter \( \beta \) is linearly related with an increase in the detrainment rate and a decrease of the erosion coefficient. So, Fig. 1 also implies that, as should be expected, cloud fraction increases with the detrainment rate and decreases with the erosion coefficient. A more detailed comparison with observations and other schemes is carried out in Teixeira (2001).

It is important to note that the relation \( l_c = \beta q_s \), that is used in this simple cumulus cloud scheme, in the determination of the cloud optical properties and in the stratocumulus diagnostic scheme, cannot be derived from first principles. This relation fails to capture some fundamental feedback mechanisms of the thermodynamics of the climate system (e.g., Betts and Harshvardan 1987). However it must be pointed out that this type of relation has been successfully used in large-scale atmospheric models (e.g., ECMWF 1991).

In order to investigate if this relation is valid at least in a statistical sense we analyzed in detail LES results from three intercomparison studies: (i) ASTEX stratocumulus (Duynkerke et al. 1999), (ii) Atlantic Tradewind Experiment (ATEX) cumulus-under-stratocumulus (Stevens et al. 2001), and (iii) Barbados Oceanographic and Meteorological Experiment (BOMEX) cumulus (Siebesma et al. 2001, manuscript submitted to J. Atmos. Sci.). From this analysis we can conclude that the linear relation between the in-cloud liquid water and the saturation specific humidity is valid in a first approximation and in a mean statistical sense (time and spatial mean). The values for the proportionality coefficient suggested by this study oscillate between 0.01 and 0.05, which is in the range of values that we will use in our simulations.

In a certain sense this relation between the in-cloud liquid water and the saturation specific humidity is as justifiable as Slingo’s (1987) cloud scheme that relates stratocumulus cloud amount to the inversion strength. It is clear that there is no obvious local and instantaneous relation between cloud amount and stability, but statistically this seems to be the case for the stratocumulus regions.

c. A stratocumulus diagnostic scheme

LES studies from Cuijpers and Bechtold (1995, hereafter CB95) have been recently used to validate statistical cloud schemes and have shown that, for boundary layer clouds, the cloud fraction can be diagnosed from a simple relation. In the present work, a simplified version of the relation from CB95 is used. It is assumed that the cloud fraction can be diagnosed as
\[ a = 0.5 + \alpha \arctan (\gamma Q), \]  
where \( \alpha = 1/\pi \), \( \gamma = 1.55 \), \( Q \) is defined as 
\[ Q = \frac{q_e - q_i}{\sigma}, \]  
and \( q_i = l + q \) is the total water content.

We assume that the standard deviation \( \sigma \) can be determined as 
\[ \sigma = \lambda \sqrt{\frac{q_i q_s}{}\eta \sqrt{\overline{\theta^2 \theta^2}}}, \]  
where \( \lambda \) and \( \eta \) are constants, \( \theta \) is the potential temperature, and the variance of potential temperature is obtained from the steady-state version (where the transport term is neglected) of the prognostic variance equation (e.g., Stull 1989):
\[ \frac{\overline{\theta^2 \theta^2}}{\tau} = -\delta \overline{\theta^2} \left( \frac{\partial \theta}{\partial z} \right) = \delta k \left( \frac{\partial \theta}{\partial z} \right)^2, \]  
where \( k \) is the turbulent diffusion coefficient, \( \tau \) is a dissipation timescale, and \( \delta \) is a constant. This type of assumption (steady state and neglecting the transport term in the variance equation) is a common procedure (e.g., Bechtold et al. 1995; Lenderink and Siebesma 2000; Siebesma and Teixeira 2001).

Last, the standard deviation is parameterized as 
\[ \sigma = B \sqrt{k} \left( \frac{\partial \theta}{\partial z} \right), \]  
where \( B \) is a constant.

The turbulent diffusion coefficient is simplified to 
\[ k = (1 - a)k_d + a k_c, \]  
where \( k_d \) and \( k_c \) are constant values and are, respectively, the turbulent diffusion coefficients for the dry and the cloudy boundary layer. By generally imposing \( k_c \gg k_d \), we are basically assuming that, when there is a cloud, turbulence is mostly produced within the cloud.

Taking this into account and assuming that \( q_i = al \) \( q = a \beta q_l \), \( q \), we obtain the following iterative equation to diagnose cloud fraction:
\[ a_{i+1} = 0.5 + \alpha \arctan \left( \frac{\gamma q_i [\beta a_i - (1 - RH)]}{B \sqrt{(1 - a_i)k_d + a_i k_c} \left( \frac{\partial \theta}{\partial z} \right)} \right). \]  

Several offline tests, for all sorts of realistic values of saturation-specific humidity, relative humidity, and inversion strength, have shown that the previous equation converges very rapidly with usually less than 10 iterations being needed in order to achieve an accurate result. An initial value of zero has been used. This equation is used at every vertical level below the PBL inversion but the vertical gradient of potential temperature considered is only the one at the PBL inversion.

To allow some insight into what this scheme is actually doing, some results from simple offline tests are shown. In these tests, the values used for the iterative Eq. (10) are \( k_d = 25 \text{ m}^2 \text{ s}^{-1} \), \( k_c = 0.25 \text{ m}^2 \text{ s}^{-1} \), and \( B = 10^{-2} \text{ s}^{1/2} \text{ K}^{-1} \). The values for the turbulent diffusion coefficients basically follow the work of van Meijgaard and van Ulden (1998) and show that the scheme is mostly attempting to parameterize the turbulent diffusion associated with the cloud itself. The saturation vapor pressure is set to \( q_i = 15 \text{ g kg}^{-1} \) and \( \beta = 0.05 \). As an illustration of the results, Fig. 2 shows the cloud fraction versus the relative humidity for different boundary layer inversion values. It can be seen that stronger inversions lead to larger values of cloud fraction for relative humidity below 0.96, and to smaller values of cloud fraction above that. This behavior is consistent with the assumption of a probability distribution for the thermodynamical variables: if far from saturation, an increased variance (stronger inversion) leads to larger cloud fraction and, if close to saturation, a larger variance implies a smaller cloud fraction. The growth of cloud fraction with stronger inversions, for most of the relative humidity values, is in agreement with the Slingo (1987) scheme and with what has been found by Klein and Hartmann (1993).

Figure 2 also confirms that for very small values of the variance, the new scheme tends asymptotically to an “all-or-nothing” cloud scheme. The sensitivity to the values used in our crude parameterization of the turbulence in the cloud layer is not shown, since it basically shows the same as for the inversion strength: larger (smaller) values of the diffusivity coefficients lead to larger (smaller) values of the variance.

It should be noted that using a more complete definition of \( \sigma \) (e.g., CB95) would provide a more accurate representation of the variance of the thermodynamical...
variables. However, taking into account the other simplifications, we consider that (6) is a reasonable approximation.

d. Implementation of the new cloud scheme in NOGAPS

The NOGAPS model uses the Slingo (1987) cloud scheme. This scheme has parameterizations for deep convective clouds, based on convective precipitation, and for boundary layer clouds, based on the boundary layer inversion strength. It also has a general cloud cover parameterization at all model levels, based on a quadratic dependency between cloud fraction and relative humidity.

The new cloud scheme is a combination of both schemes described above in sections 2b and 2c: a diagnostic cloud fraction relation for cumulus, coupled to the shallow convection scheme, and a diagnostic cloud fraction relation for PBL clouds in general, based on CB95. A relation based only on relative humidity is still used but with a relative humidity threshold value of 0.97 (in the operational scheme it is 0.9). This is done in order to try to capture a certain degree of variance of the thermodynamical properties, not taken into account by the two schemes presented in section 2b and 2c.

In the present version of the new scheme, Eq. (3) is implemented in a simplified way. The scheme uses information from the convection scheme to set the cloud base and top. Then, the detrainment and the erosion coefficients are specified to constant values: \( D = 5 \times 10^{-4} \text{ s}^{-1} \) and \( K = 5 \times 10^{-8} \text{ s}^{-1} \), for shallow convection (\( \beta = 0.01 \)). The deep convection is also parameterized with the new scheme, using Eq. (3) instead of the Slingo (1987) convective cloud scheme, with \( \beta = 0.01 \), \( D = 10^{-6} \text{ s}^{-1} \) and \( K = 2 \times 10^{-7} \text{ s}^{-1} \). The inverse timescale \( 1/K \) for deep convective clouds corresponds (with \( RH = 0.8 \) and \( \beta = 0.01 \)) to an e-folding time for the cloud dissipation of almost 3 days. We specify \( D \) and \( K \), not only because it is simpler, but also because in Teixeira (2001) it was found that it is probably more important to maintain a realistic balance between the creation of clouds due to detrainment and destruction due to the erosion, than to use a very specific value of detrainment from the convection scheme. In this way the clouds are coupled to the convection schemes but are independent of the details of each convection scheme. The values used for the iterative Eq. (10) are \( k_c = 25 \text{ m}^2 \text{s}^{-1} \), \( k_d = 0.25 \text{ m}^2 \text{s}^{-1} \) (e.g., van Meijgaard and van Ulden 1998), and \( B = 2 \times 10^{-3} \text{s}^{1/2} \text{K}^{-1} \). The liquid/ice water and the cloud optical properties are described as described at the end of section 2a.

3. The low cloud cover and the shortwave radiation at the surface

The NOGAPS model with a resolution of T63L24 was used to simulate the June–July–August (JJA) season of 1999. For each month a 40-day simulation was carried out using the initial conditions provided by the operational analysis and the daily analysis of SST and sea ice concentration. The results presented are the mean values for these 3 months (discarding the first 10 days of each simulation). This was done in order to compare with climatologies of low cloud cover and surface shortwave radiation. The low stratus cloud climatology is obtained from surface observations compiled by Warren et al. (1986, 1988). The climatology for shortwave radiation has been compiled by Da Silva et al. (1994).

In Fig. 3a, the low stratus cloud cover global distribution, according to the climatology of Warren et al. (1986, 1988) for the JJA season is shown (cloud cover units are in percent). This climatology is based on surface synoptic observations, and is probably particularly realistic for low clouds, since they can be seen well by the synoptic observer. The white areas correspond to regions where not enough data was available to produce a statistically significant climatology. The areas of the globe where low clouds are more frequent during the Northern Hemisphere (NH) summer are quite clear in this figure. These are the tropical and subtropical clouds areas off the west coast of continents, the Arctic stratus, and the fog regions off the northeast coast of continents.

In Fig. 3b, the NOGAPS low clouds for the control simulation, in the period JJA, are shown. When compared with the observations, it can be seen that the fog regions are reproduced by the model in a realistic way, but the subtropical clouds are severely underestimated. Except for fog at the subtropical west coast of continents, there are very few subtropical clouds in the control simulation. Actually over the ocean, the regions where the observations show the highest values of subtropical low cloud cover, are in the model the areas with less clouds: in some regions with no clouds whatsoever.

Other problems in the representation of low clouds in the control model are that (i) the Arctic stratus is overestimated by the model by about 20%, (ii) in the western parts of the oceanic subtropics the model produces far too much low clouds due to deep convection, and (iii) in the Indian Ocean the model produces too much spurious clouds that are again related to the deep convection cloud scheme.

The low cloud results with the new scheme are shown in Fig. 3c. The new scheme is able to reproduce the distribution of subtropical boundary layer clouds in a much more realistic way. Besides the Arctic stratus and the fog areas off the northeast coast of continents, the areas with more low cloud cover, in the new scheme, are in the subtropical regions. The actual values compare very well with the observations, with cloud cover above 50% in large areas of the subtropics.

A problem in the new scheme is that the peak value of cloud cover in the subtropical areas, is too far off the coast. However, this is also a problem in more complex cloud schemes like the one used at ECMWF (e.g., Teixeira 1999a; Duynkerke and Teixeira 2001).
FIG. 3. (a) Low stratus cloud cover global distribution, according to the climatology of Warren et al. (1986, 1988) for the JJA season; (b) low clouds from the NOGAPS control model for JJA 1999; and (c) low clouds from NOGAPS with the new cloud scheme for JJA 1999. The cloud cover units are in percent.
The new scheme produces results that are better than the old model, in several other areas of the globe. The Arctic stratus cloud cover is overestimated by the old model, while the new scheme produces realistic cloud cover in the Arctic. Also, the old model produces an excessive amount of clouds in the deep convection areas and, in general, produces a fair amount of clouds in the subtropical and tropical areas where the observations show very few low clouds.

In the fog region off the northeast coast of Asia, the
new scheme underpredicts the fog amount, while the old model overpredicts it. In the fog area off Newfoundland, the new model is quite realistic while the old model predicts an excessive amount of fog.

It should be noted that the climatology is for low stratus clouds and not for low clouds in general. In the stratocumulus areas the difference between low stratus and low cloud amounts is probably very small. In regions with convection, however, there is a difference between the two quantities. It may well be that the new scheme is slightly underpredicting the cloud amount in some deep convection areas, but it is also probable that the slight overestimation of cloud amount in the trades is only showing a lack of cumulus clouds in the climatology. Nevertheless, the difference between low clouds and low stratus is probably small enough, not to affect the general analysis of the results.

In Fig. 4a, the ocean climatology of net surface short-wave radiation (SSWR) based on the Comprehensive Ocean–Atmosphere Data Set (COADS) data (Da Silva et al. 1994), for the JJA season is shown ($W m^{-2}$). The white areas correspond to land areas or regions where the values are less than 20 $W m^{-2}$. The eastern regions of the subtropical oceans have lower values of SSWR than the western counterparts. This is due to the high prevalence of subtropical BL clouds in the eastern areas of the subtropical oceans. Relatively low values of SSWR are also present above latitude 40°N, due to the high frequency of fog and Arctic stratus.

In Fig. 4b, the NOGAPS SSWR for the control simulation, in JJA 1999, is shown. When compared with the observations, the SSWR in the stratocumulus and cumulus regions is severely overestimated (up to about 100 $W m^{-2}$ in places). There is a major difference between the western and eastern side of the tropical and subtropical oceans, that is particularly striking in the Pacific Ocean: in the climatology the eastern side has lower values of SSWR due to subtropical clouds, while in the control model the SSWR is much lower in the western side of the ocean. Such a major difference between the model and the climatology is bound to have serious consequences in terms of coupled ocean–atmosphere modeling. Also, there are in the control model extremely low values of SSWR in some tropical areas where deep convection is prevalent. These low values (down to about 60 $W m^{-2}$ in places) have no correspondence in the observations and are due to the overproduction of clouds associated with deep convection. We should not forget, however, that we are comparing JJA 1999 values for the model against a climatology, but in any case, such a large mismatch cannot be explained solely by differences between values of a single season and climatology.

The SSWR results with the new scheme are shown in Fig. 4c. There is a better agreement between the observations and the new model in the major stratocumulus and cumulus regions like off California, South America, and Namibia. In these regions the behavior of SSWR is, in some instances, quite well reproduced with the simulation of “cold tongues” of low SSWR values in the equatorial east Pacific and Atlantic. These cold tongues are, in the new scheme as in the observations, due to the presence of subtropical BL clouds, and should not be confused with reasonably similar features, in the old model, that are due to the presence of deep convection. The new scheme, however, underestimates in general the SSWR in the subtropical regions. In any case, this problem is a minor issue when compared with the deficiencies of the control model. This can also be confirmed by noting that in the new scheme, as in the observations, the values of SSWR are lower in the eastern subtropical oceans, when compared with their western counterparts. This feature is in sharp contrast with the results from the control model.

In order to better quantify the new cloud scheme’s improvements in SSWR, two major stratocumulus areas were used to compute mean values of SSWR from the two models and the climatology: the Californian stratus area ($15^\circ N, 120^\circ W, 35^\circ N, 140^\circ W$) and the Peruvian stratus area ($0^\circ, 80^\circ W, 20^\circ S, 100^\circ W$). The mean climatological SSWR for the Californian area is around 230 $W m^{-2}$, the old model’s mean SSWR is around 285 $W m^{-2}$, and the new model’s value is about 220 $W m^{-2}$. In the Peruvian region the mean climatological SSWR is around 160 $W m^{-2}$, the old model’s mean is 220 $W m^{-2}$, and the new model’s mean is about 160 $W m^{-2}$. A significant improvement with the new scheme is clear in these numbers. In both regions the new scheme reduced the old model’s positive bias by about 60 $W m^{-2}$ leading to realistic mean values of SSWR.

All this shows that the new scheme produces a more realistic distribution of SSWR in the global oceans. But the SSWR results also indicate that improving only the parameterization of cloud fraction is not enough, and that improvements are needed in the parameterization of the cloud water content. As a last note, it can be seen that the global mean values of SSWR for both experiments are very similar and are close to global observations (Gupta et al. 1999). It should also be mentioned that similar improvements were obtained for the NH winter season, and because of this reason, the results for the DJF 1999 season are not shown.

4. The transition from stratocumulus to cumulus

The evolution of stratocumulus downwind the trades and its transition to cumulus is a fundamental aspect of the dynamics of the subtropical boundary layer. The ASTEX observational program (carried out from 1–28 June 1992) was designed to study stratocumulus and the transition to trade cumulus (Albrecht et al. 1995). The ASTEX Lagrangian I (Bretherton et al. 1999) was a particular observation period where the trajectory of the boundary layer was followed for 42 h, during which it advected 1400 km from ($41^\circ N, 24^\circ W$) to ($29^\circ N, 29^\circ W$). The cloud changed from a well-mixed stratocumulus
deck to a situation with cumulus under stratocumulus and then to a 160-hPa-thick layer filled mainly with cumulus. Detailed studies of this particular observation period (e.g., DeRoode and Duynkerke 1997) have helped to understand some of the mechanisms that control the evolution and the transition of the subtropical boundary layer.

A realistic simulation and a better understanding of the complex processes involved in this transition are fundamental for more accurate climate predictions and weather forecasts. Several studies have been dedicated to this issue (e.g., Bechtold et al. 1995; Bretherton et al. 1999) and the idea of having unified PBL and cloud parameterizations capable of representing stratocumulus, cumulus, and the transition is a long-pursued goal (e.g., Bechtold et al. 1995; Teixeira and Cheinet 2002).

In order to show the vertical structure of the subtropical clouds and the transition from stratocumulus to cumulus in the new scheme, the cloud cover at 3 pressure levels (1000, 900, and 800 hPa) within the boundary layer is shown in Fig. 5. The area selected corresponds to the subtropical boundary layer clouds off the coast of California. The other corresponding areas show similar features.

It can be seen that at 1000 hPa (Fig. 5a) the clouds are concentrated close to the coast, the cloud cover values decreasing away from the coast. With the boundary layer growing down the trades, due to higher SST and lower subsidence, the clouds are located at higher levels as can be seen from Fig. 5b, where the cloud cover at 900 hPa is shown. Figure 5c shows that deep into the “trade wind” boundary layer, clouds are clearly present at 800 hPa. These results look quite reasonable and it would not be excessive to say that the new cloud scheme produces a realistic transition from stratocumulus to cumulus in terms of cloud cover.

In order to get a better insight in the way the new scheme performs in terms of this transition, a cross section of the relative humidity from the coast of California (35°N, 240°E) to the deep Tropics (5°N, 180°E) is shown in Fig. 6a. It can be seen that the model’s PBL grows along the trades with increased SST and decreased subsidence. When getting closer to the Tropics, the PBL height (e.g., as defined by a strong gradient in relative humidity), somehow ceases to exist, as deep convection kicks in. This overall behavior of the model’s PBL confirms that the model is able to represent the dynamics of the subtropical PBL. The model’s subtropical PBL height, when compared with observations and climatologies, is quite realistic, going from a shallow and foggy PBL along the California coast to a cumulus region that is about 150 hPa deep. These results are in general similar to other models (e.g., Teixeira 2000; Moeng and Stevens 2000).

In Fig. 6b, a similar cross section corresponding to the stable PBL clouds (produced by the stratocumulus scheme) is shown. The scheme is able to produce PBL clouds in the subtropics in a reasonable way. It produces fog in the coast of California that grows into stratocumulus and then almost no clouds in the deep convective regions. Unfortunately the peak value for stratocumulus is too far off the coast when compared with observations. This is even clearer in the cross section (mean of July 1999) than in the low cloud cover distribution of Fig. 3c (mean of JJA 1999). This peak in the model’s low stable cloud cover is related to the area of the highest values of relative humidity and inversion strength. So, if the highest values of relative humidity and inversion strength were closer to the coast as climatologically expected, the cloud scheme would produce a more realistic cloud cover. However, it should be noted that in some situations such as ASTEX Lagrangian I, the strongest inversions have been observed to be on the shallow convection areas.

The evolution of the PBL convective clouds is shown to be also reasonable in the cross section of Fig. 6c, following the general pattern set by the overall behavior of the PBL. The PBL convective cloud amount in the new scheme is determined by the relative humidity and by the triggering of the shallow convection scheme. So, shallow cumulus can only exist if there is enough conditional instability so as to trigger the shallow convection scheme. This explains why the shallow cumulus is only present in certain areas. As a consequence of the dependence on relative humidity the peak values of shallow convective cloud cover are located in similar areas as the peak values of stable PBL clouds.

5. Assimilation/forecast experiments and the Arctic’s PBL temperature

The new scheme was also tested in the context of data assimilation and medium-range forecasts. Two periods (one during the NH summer and one during the NH winter), of two months each, of assimilation and forecasts with the T79L24 NOGAPS system were performed. In order to study the impact of the new model, on the large-scale dynamics, with the same resolution as the operational version of NOGAPS, one month (September 1999) of T159L24 assimilation/forecast simulations was also performed.

It has been known for some time that the NOGAPS model overestimates the temperature in the Arctic boundary layer, particularly during the NH winter. Analysis of Surface Heat Budget of the Arctic (SHEBA) data (A. Beesley and R. Preller 2001, personal communication) has shown that an important cause of this warm bias is the model’s overestimation of low clouds. For the 2-m temperature, NOGAPS overestimates the observed temperature at the SHEBA locations (from December 1997 to January 1998) by an average of about 8 K. For the total cloud cover, the positive bias is on the order of 5 octas. Also, the periods when NOGAPS temperature is close to the observations are strongly correlated with the periods of accurate cloud cover prediction. Similar positive temperature biases are also pre-
Fig. 5. Cloud cover from NOGAPS with the new cloud scheme (JJA 1999) at 3 pressure levels: (a) 1000, (b) 900, and (c) 800 hPa. Units are in percent.

Figures 7a and 7b show the mean temperature error (respectively, at 1000 and 850 hPa) for the North Pole region (north of 60°N) during the NH winter T79L24 experiment (from 1 January to 28 February 2000). For the control model, the evolution of the error (defined as the forecast minus the analysis at the corresponding time) in terms of the forecast range (h) shows a growth...
from a zero error at 0-h forecast range to about 2 K (1000 hPa) and 1.4 K (850 hPa) at 120-h forecast range.
A significant improvement in the North Pole PBL temperatures is evident. The simulation with the new cloud scheme has a positive bias that is reduced, when compared to the control experiment, by more than 0.5 K (at 1000 and 850 hPa) at 120 h (5 day) forecast range.
During the NH summer there is also a boundary layer warm bias over the North Pole region. As can be seen in Figs. 7c and 7d the results with the new model, when compared with the control simulation (for the T159L24 experiment), are even better than during the NH winter.
The main reason for this substantial improvement with the new scheme, in the Arctic PBL temperature is a reduction of the Arctic cloud cover, both in summer and winter, as can be seen in Fig. 3. With the new cloud scheme the low cloud amount over the North Pole region is substantially less, and closer to the observations.

6. Seasonal and diurnal cycle
The two major modes of solar forcing variability (diurnal and seasonal cycle) induce a substantial response on the subtropical boundary layer clouds. Stratocumulus has well-defined diurnal (e.g., Duynkerke and Teixeira 2001) and seasonal oscillations (Klein and Hartmann 1993). In order to test how the new cloud scheme behaves in terms of the diurnal and seasonal cycles, the low cloud amount results for two major stratocumulus areas (off Namibia and California) are studied in detail. The low cloud cover results from the operational NOGAPS model are analyzed for the months of January and August 2001.

In Fig. 8a the low cloud cover versus the forecast range (12–120 h, starting at 0000 UTC) is shown for the Californian stratus area (15°N, 120°W, 35°N, 140°W). The low cloud cover is larger in August than in January, with a mean value of about 50% in August and about 40% in January. This result is in agreement with what is generally expected to occur (e.g., Klein and Hartmann 1993). In the NH summer the overall PBL stability (e.g., difference between the 700-hPa potential temperature and the SST) is larger than during the NH winter leading to moister (in terms of relative humidity) boundary layers and consequently more stratocumulus during the NH summer. Results from Klein and Hartmann (1993) indicate that for this region (their area is smaller) we should expect a climatological value of about 45% for the January low cloud cover and about 60% for August. Their results are actually only for full seasons, with 67% of JJA low cover, but assuming a linear low cloud cover evolution and taking into account the decrease on stability from June to August, it is probably reasonable to assume a 60% low cloud cover in August. Comparing the model results with the climatology, and taking into account that the model results are for a particular August and January, it can be said that the model’s low cloud cover is quite realistic (al-
though the model still probably underestimates the NH summer low cloud cover).

Also present in Fig. 8a is a noticeable diurnal cycle. Since the forecast starts at 0000 UTC, night in this region corresponds to 12 h, 36 h, and so on. The model has a diurnal oscillation of about 5% in January and about 10% in August. Duynkerke and Teixeira (2001) show a substantial diurnal cycle of stratocumulus off California of about 10%–20% in July’s cloud cover from the First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment (FIRE) data, which makes the model value of about 10% for August quite realistic. The larger diurnal oscillation during August is also a realistic model representation of the more intense diurnal solar forcing during the NH summer.

In Fig. 8b the same is shown but for the Namibian stratus area (10°S, 10°W, 30°S, 10°E). Again the model shows a substantial seasonal difference between the low cloud cover for January and August 2001. In August the mean cloud cover is about 47% and in January is about 37%. The fact that the Namibian stratus cloud cover is higher in August is quite realistic and is again connected by Klein and Hartmann (1993) to a higher PBL stability in the NH summer when compared with the NH winter. When compared with the seasonal climatology (about 50% for DJF and 65% for JJA) the model seems to underpredict the low cloud cover, although not by a substantial amount. It is also interesting to note that although the January mean cloud cover values are smaller, its diurnal oscillation is much larger than in August (an oscillation of more than 10% in January as compared with about 3% in August). Again this is a consequence of a much stronger solar cycle during January (SH summer).

7. NOGAPS vertical resolution and the representation of PBL clouds

A realistic representation of the boundary layer processes requires an adequate vertical resolution in the lower troposphere of global atmospheric models. The impact of vertical resolution in boundary layer processes in the context of global models has been the subject of some recent papers (e.g., Beljaars 1991; Delage 1988).
More specifically regarding the cloud-topped boundary layer, there have been some studies that suggest that the current vertical resolution of NWP and climate models is too low for an accurate representation of the boundary layer physics. Van Meijgaard and van Ulden (1998) developed successfully a one-dimensional cloudy boundary layer model and tested the sensitivity of their model to different vertical resolutions. They found that there is a limit on the vertical grid spacing of about 200 m, above which the cloud-top physics and dynamics cannot be simulated properly. Lenderink and Holtslag (2000) produced similar findings in their 1D simulations.

In global atmospheric models the vertical resolution is usually not uniform. The finest resolution in height is near the surface and degrades upward. In terms of pressure, the resolution is also higher close to the surface and degrades upward until some level in the troposphere, above which it improves again. This general behavior, from the surface to 500 hPa can actually be seen in Fig. 9, where the thickness (in pressure) of each of the levels versus the pressure is shown for 4 different vertical configurations. These are the NOGAPS operational configuration of 24 levels, the new NOGAPS vertical resolution of 36 levels that is currently being tested (T. Hogan 2001, personal communication), the 31 levels configuration of ECMWF that was operational until October 1999, and the one recently implemented at ECMWF with a vertical resolution of 60 levels (Teixeira et al. 1999; Jakob et al. 2000).

In Fig. 9 it is shown that in the NOGAPS L24 resolution used in this study, only the first three levels have a resolution better than 20 hPa. However, the L24 configuration is fairly reasonable when compared with the ECMWF L31 configuration in the boundary layer (both configurations have six levels below 850 hPa). The new NOGAPS L36 resolution is a substantial step ahead, being clearly better in the PBL than the ECMWF L31 resolution and fairly competitive with the new ECMWF L60 configuration. It must be noted, that all of the vertical configurations shown in Fig. 9 can be considered high-resolution configurations if compared with other global NWP or climate models, and that the ECMWF L60 configuration is the best resolution used in global NWP models around the world. So, in conclusion it can be said that although the NOGAPS L24 PBL resolution is not yet good enough to fully represent the PBL dynamics, it is in general quite superior to the resolutions usually used in most climate models (e.g., Slingo et al. 1995).

It is also clear that the simple cloud cover parameterizations described in this study will not produce a perfectly convergent scheme for very high vertical res-
olutions. Unfortunately this is true for a substantial part of all sorts of parameterization schemes used in NWP and climate models. With the exception of most radiation and vertical diffusion schemes, all other types of schemes like moist convection (e.g., Tiedtke 1989), clouds or gravity wave drag often suffer from severe convergence problems (see Teixeira 1999b for some details). Even radiation and vertical diffusion schemes are rarely tested for numerical convergence. Nevertheless, recent experimentation with the new NOGAPS L36 resolution shows that the new cloud scheme is quite robust in terms of its sensitivity to changes in the vertical configuration (at least from L24 to L36).

8. Summary

Boundary layer clouds have a major impact on the surface weather and climate. In global atmospheric models these clouds are usually not realistically represented. In this paper, we propose two simple cloud cover prediction schemes for cumulus and stratocumulus. One is based on a steady-state version of the prognostic cloud fraction equation suggested by Tiedtke (1993). The other is based on a simplified version of the cloud scheme suggested by LES studies from Cuijpers and Bechtold (1995). These two formulations are combined as a new cloud scheme and implemented into the NOGAPS model.

The NOGAPS model, with the old and the new cloud scheme, was used to simulate the June–July–August (JJA) season of 1999. The results for low clouds were compared to a climatology obtained from surface observations. In the old cloud scheme the subtropical clouds are severely underestimated. The new scheme, on the other hand, is able to reproduce the distribution of subtropical boundary layer clouds in a much more realistic way. The actual values compare well with the observations, with cloud cover above 50% in large areas of the subtropics.

A comparison with climatological values of net surface shortwave radiation shows that the new scheme is also clearly able to improve the distribution of the shortwave radiation at the ocean surface in the Tropics and subtropics. This result can have a large positive impact on coupled ocean–atmosphere modeling.

It is also shown that the vertical structure of the subtropical PBL and the transition from stratocumulus to cumulus is well reproduced in the new scheme. In the context of data assimilation and medium-range forecasts, it is shown that the new scheme produces a significant improvement in the Arctic PBL temperature, by reducing a severe warm bias due to excessive cloud amount in the old model. It is also shown that the new scheme is able to realistically reproduce the seasonal and diurnal cycles of stratocumulus cloud cover.

As a last note, it should be mentioned that the cloud scheme presented in this paper has been implemented operationally, in December 2000, in the U.S. Navy’s NOGAPS global weather prediction system (Teixeira and Hogan 2001) at Fleet Numerical Meteorology and Oceanography Center. In terms of the overall scores for geopotential in the Northern and Southern Hemispheres the new model is slightly better.

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