THREE-DIMENSIONAL SIMULATION OF CLOUD
STREET DEVELOPMENT DURING A COLD
AIR OUTBREAK

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Abstract. A three-dimensional numerical model is used to study boundary-layer eddy structure during a cold air outbreak. The model explicitly represents the large-scale three-dimensional motions, while small-scale turbulence is parameterized; it contains a water cycle with cloud formation and it takes into account infrared radiative cooling in cloudy conditions and the influence of large-scale vertical motions.

The model is applied to conditions corresponding to an observed case of cloud street/stratocumulus development which occurred over the Greenland Sea during the ARKTIS 1988 experiment. The boundary layer is found to grow rapidly as the cold air flows off the ice over the relatively warm water. Coherent structures were identified in this boundary layer. It is found that the rolls become increasingly more convective in character with distance from the ice edge. Qualitative and quantitative descriptions of the flow field are given. Additionally, the relative importance of the various physical processes and external parameters in the evolution of the mean field of variables is indicated.

1. Introduction

Convective activity is usually associated with cold air outbreaks over warm water. In the case of the Greenland sea, northerly winds bring cold and dry polar air masses from the ice pack over the relatively warm open waters of the ocean. Cloud streets formed 100 km or less downstream from the coast line provide flow visualization of the flow patterns developing within the convective boundary layer. The conception is that the secondary flow consists of horizontal roll vortices which extend throughout the depth of the boundary layer. Their axis is roughly in the direction of the basic flow and parallel stripes of upward motion within these helical motions may in some cases be marked with clouds. In other cases, the rolls are indicated by rows of dense dark clouds within a solid cloud cover. The spacing between adjacent cloud lines is normally between 2–8 km and they persist for several hundred kilometers south of the ice edge where a transformation takes place to a three-dimensional cellular convective regime.

The conditions for the formation of these rolls have been the subject of many theoretical and observational studies, e.g., Küttner (1971), Le Mone (1973) and Brown (1980). The cause of the rolls has been attributed to both a shear instability of the Ekman planetary boundary layer (PBL) and the organization of buoyant convection by velocity shear. Often it is a combination of these two instability mechanisms. A number of observational studies of cloud street formation during cold air outbreaks (Walter, 1980; Miura, 1986) have provided information as to
how the interval between the cloud lines changes with distance south of the edge of the ice pack, but not on turbulent and roll-scale transports of momentum and energy. Therefore, it might be useful to apply a numerical model to the cold air outbreak problem, with special emphasis on boundary-layer eddy structure and turbulent statistics.

The observed two-dimensionality has formed the basis for several numerical studies, e.g. Mason and Sykes (1982), Mason (1985), Etling and Raasch (1987), Chlond (1987), Sykes et al. (1988, 1990), and Raasch (1990). These studies have provided a better understanding of the physical processes involved, and have given detailed information on roll-scale velocity variances and roll-scale vertical transports. However, since the models used have assumed homogeneous conditions in the direction of the roll axis, they cannot predict whether two-dimensional eddies should actually dominate the flow. In addition, calculations by Chlond (1987) lead to the suggestion that the relation between the cross-roll and along-roll velocity variances cannot be reproduced correctly by two-dimensional models.

Therefore, we shall try to extend this work. Unlike previous studies of the moist convective boundary layer, which were subject to limitations in the ability to resolve the actual eddy structure, our primary interest is in simulating the temporal development of three-dimensional boundary-layer flows during cold air outbreaks under conditions where the surface heat flux, latent heat release and radiative fluxes should have a strong effect. Hence, a three-dimensional numerical model has been developed in order to take into account circulations acting in both the horizontal and the vertical planes. The general idea underlying the model is that of a large-eddy model. The model explicitly calculates the spatial averages, which hopefully represent the dominant large-scale motions, while parameterizing the effect of the fluctuations on the averaged flow quantities. The model includes most of the physical processes occurring in a moist boundary layer in the absence of precipitation. It contains a water cycle with cloud formation (including a subgrid-scale condensation scheme), a treatment of the subgrid scale turbulence which incorporates effects of thermal stratification; it takes into account infrared radiative cooling in cloudy conditions (using an effective emissivity model) and the influence of large-scale vertical motions. The model is applied to conditions corresponding to an observed case of cloud street stratocumulus development which occurred over the Greenland Sea during the ARKTIS 1988 experiment (Brümmer et al., 1991). Our principal objectives are to determine the respective roles of condensation, cloud top radiative cooling, large-scale subsidence, the large-scale moisture field and time-dependent surface heating on boundary-layer rolls by considering simulations with and without these effects. Although the model is far from perfect for the simulation of such complex boundary-layer phenomena, it can give at least some insight into the exchange process between ocean and atmosphere due to organized vortices in the boundary layer.
2. Model

The basic dynamic framework of our model follows that of a large-eddy simulation (LES) model. For three-dimensional turbulence, this type of model must be time dependent and should have a grid size which falls within the inertial subrange. The LES approach was first applied to meteorological flows by Deardorff (1972) and Sommeria (1976); it is discussed and reviewed by Herring (1979) and Schumann and Friedrich (1986).

2.1. Governing equations

The fluid is supposed to consist of a mixture of two perfect gases, air and water vapour, containing water droplets in suspension. The equations governing the resolved-scale motions are the three-dimensional, spatially averaged incompressible, Boussinesq equations conserving mass, momentum, liquid water potential temperature and total moisture content. The calculations are carried out in a Cartesian coordinate system, whose y-axis is aligned parallel to the geostrophic wind vector. At time \( t = 0 \), the model domain is located on the shoreline and is then translated in the y-direction at a constant velocity (equal to the geostrophic wind speed). The equations then are

\[
\begin{align*}
\frac{\partial \tilde{u}}{\partial x} + \frac{\partial \tilde{v}}{\partial y} + \frac{\partial \tilde{w}}{\partial z} &= 0, \\
\frac{\partial \tilde{u}}{\partial t} &= -\frac{\partial}{\partial x} (\tilde{u} \tilde{u}) - \frac{\partial}{\partial y} (\tilde{v} \tilde{u}) - \frac{\partial}{\partial z} (\tilde{w} \tilde{u}) - \frac{1}{\rho_0} \frac{\partial \tilde{p}}{\partial x} - \frac{1}{\rho_0} \frac{\partial P_s}{\partial x} + \\
&\quad + f(\tilde{v} + v_{geo}) + \frac{\partial}{\partial x} \tau_{11} - \frac{\partial}{\partial y} \tau_{12} - \frac{\partial}{\partial z} \tau_{13} + \left( \frac{\partial \tilde{u}}{\partial t} \right)_{LS} \\
\frac{\partial \tilde{v}}{\partial t} &= \frac{\partial}{\partial x} (\tilde{u} \tilde{v}) - \frac{\partial}{\partial y} (\tilde{v} \tilde{v}) - \frac{\partial}{\partial z} (\tilde{w} \tilde{v}) - \frac{1}{\rho_0} \frac{\partial \tilde{p}}{\partial y} + \\
&\quad - f \tilde{u} - \frac{\partial}{\partial x} \tau_{21} - \frac{\partial}{\partial y} \tau_{22} - \frac{\partial}{\partial z} \tau_{23} + \left( \frac{\partial \tilde{v}}{\partial t} \right)_{LS} \\
\frac{\partial \tilde{w}}{\partial t} &= -\frac{\partial}{\partial x} (\tilde{u} \tilde{w}) - \frac{\partial}{\partial y} (\tilde{v} \tilde{w}) - \frac{\partial}{\partial z} (\tilde{w} \tilde{w}) - \frac{1}{\rho_0} \frac{\partial \tilde{p}}{\partial z} + \\
&\quad + \frac{g}{\Theta_x} (\Theta_v - \langle \Theta_v \rangle) - \frac{\partial}{\partial x} \tau_{31} - \frac{\partial}{\partial y} \tau_{32} - \frac{\partial}{\partial z} \tau_{33} + \left( \frac{\partial \tilde{w}}{\partial t} \right)_{LS} \\
\frac{\partial \Theta_t}{\partial t} &= \frac{\partial}{\partial x} (\tilde{u} \Theta_t) - \frac{\partial}{\partial y} (\tilde{v} \Theta_t) - \frac{\partial}{\partial z} (\tilde{w} \Theta_t) - \frac{\partial}{\partial x} H_1 -
\end{align*}
\]
\[ - \frac{\partial}{\partial y} H_2 - \frac{\partial}{\partial z} H_3 + \left( \frac{\partial \Theta_i}{\partial t} \right)_R + \left( \frac{\partial \Theta_l}{\partial t} \right)_{LS} \]  

\[ \frac{\partial q}{\partial t} = - \frac{\partial}{\partial x} (\bar{u}q) - \frac{\partial}{\partial y} (\bar{v}q) - \frac{\partial}{\partial z} (\bar{w}q) - \frac{\partial}{\partial x} Q_1 - \]  

\[ - \frac{\partial}{\partial y} Q_2 - \frac{\partial}{\partial z} Q_3 + \left( \frac{\partial q}{\partial t} \right)_{LS} \]  

where overbars signify the Reynolds average over the grid volume \( \Delta x \cdot \Delta y \cdot \Delta z \), and the horizontal average is denoted by the angular brackets, \( \langle \rangle \). Here, \( u, v \) and \( w \) are the velocity components, \( p \) is the perturbation pressure, \( f \) is the Coriolis parameter, \( g/\Theta_{00} \) is the buoyancy parameter and \( \rho_0 \) is the reference density, which is assumed to be independent of height. \( \tau_{ij} \) is the turbulent Reynolds stress tensor, \( H_i \) is the turbulent flux of \( \Theta_i \) and \( Q_i \) represents the turbulent total moisture flux. These fluxes are derived from the parameterization which is discussed in Section 2.2. \( P_s \) is a linearly varying background pressure so that \( \partial P_s/\partial x \) is a constant related to the \( \nu \)-component of the geostrophic wind vector by:

\[ \frac{1}{\rho_0 f} \frac{\partial P_s}{\partial x} = \nu_g . \]

In order to describe the thermodynamic properties of the cloudy PBL, and following Betts (1973) and Sommeria and Deardorff (1977), the thermodynamics are represented by conservation equations for liquid water potential temperature \( \Theta_l \) and total moisture content \( q \), instead of the traditional variables potential temperature \( \Theta \), specific humidity \( q_v \), and liquid-water content \( q_l \). These variables are defined as follows:

\[ \Theta_l = \Theta - \frac{L}{c_p} \left( \frac{\Theta}{T} \right) q_l \]  

\[ q = q_v + q_l , \]  

where \( T \) is the actual temperature, \( L \) the latent heat of evaporation and \( c_p \) the specific heat of dry air. We further assume that

\[ \frac{\Theta}{T} = \left( \frac{p_0(0)}{p_0(z)} \right)^{\kappa} = e^{\kappa \cdot z/D} , \]

where \( \kappa = 0.285 \), \( D \) is the scale height of the atmosphere set equal to 8 km, and \( p_0 \) denotes hydrostatic pressure.

In the buoyancy term, \( \Theta_v \) is the virtual potential temperature (see e.g., Sommeria and Deardorff (1977)) defined by
The liquid water content is calculated using the partial cloudiness formulation of Sommeria and Deardorff (1977), described below.

Each prognostic equation contains, on the right hand side, a large-scale corrective term (index LS) describing the effect of the inhomogeneity of the synoptic-scale meteorological fields. The possibility of considering large-scale effects (i.e., mean subsidence) is useful because it gives more flexibility to the model in order to treat actual meteorological cases and allows the existence of quasi long-term steady states with balanced budgets of momentum, heat and water.

On the right-hand side of Equation (5), \( (\partial \Theta_l/\partial t)_R \) is a cooling term simulating radiative cooling. The present study is intended to be highly idealized. Therefore, a simple treatment has been chosen, which only takes into account infrared absorption and emission by cloud droplets. Radiative cooling due to gaseous absorbers and solar radiative heating are excluded. The long-wave scheme is based on the parameterization of cloud ‘effective emissivity’. This quantity allows the calculation of broadband infrared fluxes and infrared cooling rates within the cloud. The concept of effective emissivity is not new and has been applied extensively to the cloudy atmosphere (e.g., Cox, 1976). For the flux computations at a given level \( z \), the atmosphere is supposed to be isothermal above and below \( z \), except at the surface which is considered at its actual temperature. Vertical columns are considered independent and any lateral effect is neglected. Upward and downward radiative fluxes, \( F \uparrow \) and \( F \downarrow \) are written

\[
F \uparrow (z) = B(o) + \epsilon \uparrow (z,o)(B(z) - B(o))
\]

\[
F \downarrow (z) = F(z_T) + \epsilon(z,z_T)(B(z) - F \downarrow (z_T)) ,
\]

where \( B(o) \) and \( B(z) \) are the blackbody emissions, at the temperature of the ground and level \( z \), respectively; \( F \downarrow (z_T) \) is the impinging flux at the top of the model domain \( (z = z_T) \) which has to be prescribed. \( \epsilon \uparrow (z,o) \) and \( \epsilon \downarrow (z,z_T) \) are the effective emissivities of liquid water between the ground and level \( z \), and between level \( z \) and the top of the model domain \( (z = z_T) \), respectively. They are given by

\[
\epsilon \uparrow (z,o) = 1 - \exp(-a_0 \uparrow \cdot r_l(o,z))
\]

\[
\epsilon \downarrow (z,z_T) = 1 - \exp(-a_0 \downarrow \cdot r_l(z,z_T)) ,
\]

where \( r_l(o,z) \) and \( r_l(z,z_T) \) are the amounts of liquid water contained in the columns below and above level \( z \), respectively. From Equations (11) and (12), \( a_0 \) defines a mass absorption coefficient \((\text{m}^2 \text{kg}^{-1})\) which is given by Stephens (1978) as

\[
a_0 \downarrow = 158 , a_0 \uparrow = 130 .
\]
The radiation computation is made at each time step for each column of air. The source term in Equation (5) can now be written as

$$\left( \frac{\partial \Theta_i}{\partial t} \right)_R = -\left( \frac{\Theta_i}{T} \right) \frac{1}{\rho_0 c_p \Delta z} (\Delta F(z^+) - \Delta F(z^-)) ,$$

where $\Delta F$ is the difference between upward and downward fluxes at two consecutive levels around the level where $\Theta_i$ is defined.

2.2. Subgrid parameterization

The spatially averaged equations discussed above can be solved once the subgrid-scale (SGS) fluxes are determined in terms of the resolved-scale fields. Our subgrid flux model is based on a transport equation for the SGS kinetic energy

$$\delta \equiv \frac{1}{2} (u'^2 + v'^2 + w'^2) ,$$

and is similar to that of Deardorff (1980), and Sykes et al. (1988). We write the SGS equation as follows

$$\frac{\partial \delta}{\partial t} + \frac{\partial}{\partial x_j} (u_j \delta) = -u'_i u'_j \frac{\partial \delta}{\partial x_j} + \frac{g}{\Theta_x} \delta \Theta' -$$

$$- \frac{\partial}{\partial x_j} \left( \frac{u'_j \delta'}{\rho_0} + \frac{u'_j \Theta'}{\rho_0} \right) - \frac{2\sqrt{2}}{\Lambda} \cdot b \delta^{3/2} ,$$

where the summation convention has been employed for conciseness (wherein $x_i = (x, y, z)$ and $u_i = (u, v, w)$). $\Lambda$ is a length scale specified algebraically below and $b = 0.125$ is an empirical constant for which we use the value suggested by Lewellen (1977).

The closure modeling used for Equation (15) is similar to that of Sykes et al. (1988). The subgrid fluxes were parameterized by

$$\tau_{ij} = \frac{u'_i u'_j}{3} - \frac{2}{3} \delta_{ij} \delta = K_H \frac{\partial \delta}{\partial x_j} ,$$

$$H_i = u'_i \Theta'_i = -K_H \frac{\partial \Theta_i}{\partial x_i} ,$$

$$Q_i = \frac{u'_i q'}{\rho_0} = -K_H \frac{\partial q}{\partial x_i} ,$$

$$\frac{u'_i \delta'}{\rho_0} + \frac{u'_i \Theta'}{\rho_0} = -K_H \frac{\partial \delta}{\partial x_i} ,$$

$$w' \Theta'_v = K_1 w' \Theta'_v + K_2 w' q' ,$$
where $K_M$ is a subgrid-scale eddy coefficient for momentum, $K_H$ is a subgrid-scale eddy coefficient for scalar quantities. $K_1, K_2$ are coefficients which are given by

$$K_1 = 1 - \frac{r \cdot \alpha[L/c_p - 1.61 \cdot \Theta_\infty \cdot e^{-K_z/D}]}{1 + \alpha L/c_p}$$  

$$K_2 = 0.61 \cdot \Theta_\infty + \frac{r \cdot [L/c_p \cdot e^{+K_z/D} - 1.61 \cdot \Theta_\infty]}{1 + \alpha L/c_p}$$

where

$$\alpha \equiv \left(\frac{\partial q_s}{\partial T}\right)_t = 0.622 \cdot L \cdot q_s(\tilde{T}_t)/(R_d \tilde{T}_t^2) .$$  

Here $R_d$ is the gas constant for dry air, $q_s$ is the saturation specific humidity which is obtained from the relations given by Bougeault (1981) and

$$\tilde{T}_t = \left(\frac{T}{\Theta}\right) \cdot \Theta_t .$$

The expressions (17a) – (17b) are simplified versions of those derived in Sommeria and Deardorff (1977), where we have neglected $q_s$ in comparison with unity and we used the reference temperature $\Theta_\infty$ in place of $\Theta_i$. Here $r$ denotes the subgrid-scale cloud fraction given below.

The eddy coefficients are proportional to the product of the length scale and the velocity scale $(2\tilde{e})^{1/2}$. Accordingly, we have

$$K_M = S_M \cdot A \cdot (2\tilde{e})^{1/2}$$  

$$K_H = S_H \cdot A \cdot (2\tilde{e})^{1/2}$$

where $S_M$ and $S_H$ are stability-dependent coefficients, similar to those of Mellor and Yamada (1974). Here $S_H$ and $S_M$ are given algebraically by

$$S_H = \frac{(1 - 2b)/3}{A + R_i(2 + 1/(bs))}$$

$$S_M = \frac{A^2 + (A/(bs) - 1) R_i}{A + R_i} \cdot S_H ,$$

where the turbulent Richardson number is defined

$$R_i = \frac{g}{\Theta_\infty} \left( K_i \frac{\partial \tilde{T}_t}{\partial z} + K_z \frac{\partial \tilde{q}}{\partial z} \right) \frac{\Lambda^2}{2 \tilde{e}} .$$

The constants in Equation (19) are empirical (Lewellen, 1977). We use the values $A = 1$, $s = 1.8$ and as already specified, $b = 0.125$.

The length scale $\Lambda$ is given by
\[ \Lambda = \min \left( 0.67 \cdot z \cdot \Delta, (2 \Delta)^{1/2}/N \right), \quad (21) \]

where \( \Delta \) is the average grid scale \( \Delta = 1/3 (\Delta x + \Delta y + \Delta z) \), where \( \Delta x, \Delta y \) and \( \Delta z \) are the respective grid intervals and \( N \) is the Brunt-Väisälä frequency, defined by

\[ N^2 = \frac{g}{\Theta_{00}} \left( K_1 \frac{\partial \bar{T}}{\partial z} + K_2 \frac{\partial \bar{q}}{\partial z} \right). \quad (22) \]

Hence, \( \Lambda \) is proportional to \( z \) near the surface, but is limited to the average mesh spacing. The third limit is applied only in stable regions where \( N^2 > 0 \).

The liquid water content is calculated using the formulation of Sommeria and Deardorff (1977). This formulation by-passes the shortcomings of previous condensation schemes which assume that a computational grid volume is either entirely saturated or entirely unsaturated.

This is achieved by making allowance for statistical variations of \( \Theta \) and \( q \) on the subgrid-scale. In this way, local condensation can occur in the presence of large turbulent fluctuations, even though the average state is below saturation. To derive the dependencies of mean cloud fraction and mean liquid water content upon temperature and humidity statistics, it is assumed that the instant value of the quantity \( \Delta \left( q - q_s \right) (T_s) \) (saturation deficit) is distributed around its grid volume mean value according to a Gaussian probability density function with standard deviation

\[ \sigma = \left( q_r^2 - 2 \left( \frac{T}{\Theta} \right) \alpha q \Theta \bar{T} + \left( \frac{T}{\Theta} \right)^2 \alpha^2 \Theta \bar{T}^2 \right)^{1/2}, \quad (23) \]

where \( q_r^2, \Theta_r^2 \) and \( q \Theta_r \) are the variances of \( q \) and \( \Theta \) and their covariance which are calculated diagnostically from the subgrid fluxes of \( q \) and \( \Theta \) (see below). The mean liquid water content \( \bar{q}_l \) and the fraction of grid volume \( r \) in which it is contained is then calculated by integration over every possible saturation deficit contributing to liquid water, using that probability density function. Following the usual approximations (Sommeria and Deardorff, 1977), we have

\[ \frac{\bar{q}_l}{\gamma \cdot \sigma} = \begin{cases} 
0 & Q \leq -1.6 \\
(Q + 1.6)^2/6.4 & |Q| < 1.6 \\
Q & Q \geq 1.6 
\end{cases} \quad (24a) \]

\[ r = 0.5 \left( 1 + Q/1.6 \right) \quad 0 \leq r \leq 1 \quad (24b) \]

where \( \gamma = (1 + L/c_p \alpha)^{-1} \) and \( Q \) is a normalized departure from mean saturation.
The standard deviation $\sigma$ of the saturation deficit will be needed in this calculation. Instead of solving prognostic equations for the variances of $q$ and $\Theta_i$ and their covariance, a simple diagnostic estimate (but which should have the correct magnitude) of these quantities, according to

$$Q = \frac{(\bar{q} - q_s(\bar{T}_s))}{\sigma}. \quad (25)$$

was employed.

### 2.3. Boundary and Initial Conditions

The computational domain extends horizontally and vertically over a finite domain of size $L_x \times L_y \times L_z$ (where $L_x = L_y = 6400$ m and $L_z = 1600$ m are used for all runs). At the lateral boundaries, periodicity is assumed. At the top, a zero slope condition on all variables, except $\bar{w}$, $\bar{q}$ and $\Theta_i$, is imposed. Vertical velocity $\bar{w}$ is zero at the top and the gradients of $\bar{q}$ and $\Theta_i$ are fixed at the initial gradients $\Gamma_q$ and $\Gamma_{\Theta_i}$, respectively. The upper boundary conditions do not allow the transmission of gravity waves which can be generated in a stable layer. This could lead to reflections and an unrealistic distribution of momentum and energy. To avoid reflections from the top lid, a Rayleigh friction term is added to Equation (5) and (6). In this way, a radiation condition is simulated by an artificial sponge layer in the upper third of the model domain. The coefficient of Rayleigh damping $\nu$ is gradually increased in the sponge layer according to the square of a sine function. That is

$$\nu = \nu_0 \cdot \sin^2\left(\frac{\pi}{2} \frac{(z - z_s)}{(L_z - z_s)}\right), \quad (27)$$

where $z_s$ is the base of the sponge layer and $\nu_0 = (300$ s$)^{-1}$.

At the lower boundary, we use Monin–Obukhov similarity to relate the fluxes of various quantities to the corresponding difference between the surface value and the value at the first model grid point above the surface at $z = z_1$. Using the scaling parameters $u_*, \Theta_*$ and $q_*$, we obtain the relations

$$\left(\tau_{13}^2 + \tau_{23}^2\right)^{1/2} = u_*^2 \quad (28a)$$

$$H_3 = -u_* \cdot \Theta_* \quad (28b)$$
\[ Q_3 = -u \cdot q * . \] (28c)

The scaling parameters are related to mean field variables through the following,

\[ u_* = \frac{\kappa (\overline{u}^2(z_1) + \overline{v}^2(z_1))^{1/2}}{\ln z_1/z_0 - \Psi_M(z_1/L)} \] (29a)

\[ \Theta_* = \frac{\kappa (\overline{\Theta}_T(z_1) - \overline{\Theta}_T(z_T))}{\ln z_1/z_T - \Psi_H(z_1/z_T)} . \] (29b)

A similar relation to (29b) for \( q_* \) can be written, with \( \Theta_T(z_1) - \overline{\Theta}_T(z_T) \) replaced by \( \overline{q}(z_1) - \overline{q}(z_T) \). In the above, \( \kappa \) is von Karman’s constant, \( z_0 \) and \( z_T \) are aerodynamic roughness and temperature surface scaling heights, respectively, and \( L \) is the Monin-Obukhov stability length defined by

\[ L = \frac{u_*^2}{\kappa (g/\Theta_{oo})(\Theta_* + 0.61 \cdot \Theta_{oo} q_*)} . \] (30)

The functions in Equation (29) can be related to gradient stability functions by

\[ \Psi = \int (1 - \Phi(z/L)) \, d(\ln z/L) . \] (31)

For \( z/L \leq 0 \), the functions take the form

\[ \Phi_M = (1 - \gamma_1 z/L)^{-1/4} \] (32a)

\[ \Phi_H = (1 - \gamma_2 z/L)^{-1/2} , \] (32b)

where we choose \( \gamma_1 = 20.3 \) and \( \gamma_2 = 12.2 \) according to the recommendations of Webb (1982). In addition, \( \kappa = 0.4 \) is used. Since we are interested in the marine boundary layer in particular, the liquid water potential temperature at the surface was assumed to be horizontally homogeneous but a prescribed function of time; \( \overline{q}(z = z_T) \) is set equal to the saturation value at the ocean surface. Finally, zero slope conditions are applied for \( \theta \) and \( \rho \) at the surface.

The initialization of each run proceeds in the following manner. The initial \( \Theta_T \) profile is specified by a prescribed temperature difference \( \Delta \Theta \) between the ocean surface and the overlying air and a constant gradient \( \Gamma_{\Theta} \) above. Likewise, the initial humidity profile is specified by a humidity jump \( \Delta q_T \) and a constant gradient \( \Gamma_q \) above. With these values, the initial wind profile is obtained by running a one-dimensional version of the model to steady state. During this phase, temperature and humidity profiles are fixed. In order to drive the system, this horizontally homogeneous solution is then transferred to the three-dimensional domain and perturbed by imposing small random perturbations on the temperature field at the first time step.
2.4. NUMERICAL TECHNIQUES

The numerical integration scheme is based on an equidistant staggered grid and finite difference approximations. All scalars are defined at the cell centres, $\bar{u}$ is evaluated at the middle of the upstream and downstream faces, $\bar{v}$ at the middle of the two cross-stream faces and $\bar{w}$ at the middle of the bottom and top faces. For the advection of SGS-kinetic energy we choose the upstream scheme because it guarantees that positive scalars stay positive. Otherwise, spatial differentials are approximated by second-order accurate central differences in a form which conserves the integral of linear and quadratic quantities up to very small errors (e.g., Piacsek and Williams, 1970; Williams, 1969). Euler time differencing is used in the equation for the SGS kinetic energy; in the balance equations for temperature and humidity, time integration is performed using the Adams-Bashforth scheme. The momentum equations are solved by using a predictor-corrector scheme which casts Equations (2) to (4) into

\begin{align*}
\bar{u}_i^{(N+1)*} &= \bar{u}_i^N + \Delta t \left( \frac{R_i^N - 1}{2} \frac{R_i^{N-1}}{\rho_0} - \frac{1}{\rho_0} \frac{\partial \bar{p}^N}{\partial x_i} \right) \\
\bar{u}_i^{(N+1)} &= \bar{u}_i^{(N+1)*} - \Delta t \left( \frac{1}{\rho_0} \frac{\partial \Delta \mu}{\partial x_i} \right),
\end{align*}

(33a) (33b)

where $R_i$ denotes the right-hand sides of Equations (2) to (4) except the pressure gradients and

$$\Delta \rho = \bar{p}^{N+1} - \bar{p}^N.$$  

(34)

The incompressibility condition requires the implicit determination of pressure, so that the velocity field resulting from Equation (33b) has to satisfy Equation (1). Applying the divergence operator on Equation (33b) yields a Poisson equation for the pressure increment

$$\frac{\partial^2 \Delta p}{\partial x_i \partial x_i} = \frac{\rho_0}{\Delta t} \frac{\partial \bar{u}^{(N+1)*}}{\partial x_i}.$$ 

(35)

This equation is solved by successive over-relaxation using RED/BLACK decomposition to guarantee vectorization. In this manner, the continuity equation is strictly enforced. For the runs presented here, the model domain extends to a height of 1600 m and the length treated in the horizontal directions is 6400 m. The region is split into 64 grid intervals in $x$, 64 along $y$ and 34 along the vertical $z$-direction (i.e., $\Delta x = \Delta y = 100$ m, $\Delta z = 50$ m). A time step of $\Delta t = 3$ s was used for all runs. Each simulation hour required about one computer hour on the CRAY-2S computer.
3. Results

3.1. Cases treated

A series of simulations was carried out in order to demonstrate the model's ability to serve as a tool to interpret experimental data and to understand the various physical processes acting in a cloudy convective boundary layer during a cold-air outbreak.

First, the model was applied to conditions corresponding to an observed case of cloud street/stratocumulus development which occurred over the Greenland Sea during the ARKTIS 1988 experiment. The experiment took place in the Fram Strait in the area straddling the ice margin west of Spitsbergen during the period from 5 to 27 May 1988. The experiment was dedicated to the study of boundary-layer modification and certain cloud structures in cases of off-ice and on-ice air flows.

During the experiment, four research aircraft operated from the airfield Longyearbyen on Spitsbergen: a FALCON-20, a DORNIER-128 and two DORNIER-228. FALCON-20 and DORNIER-128 were equipped with gust probes to measure mean and turbulent meteorological quantities and with a radiation thermometer to measure the surface temperature. One of the two DORNIER-228 (D-CALM) was equipped with a downward facing LIDAR system to measure cloud top heights. In the same area, surface and radiosonde observations were taken at two research vessels. Figure 1 shows the positions of the ships and the aircraft tracks on 15 and 16 May 1988. The FALCON measurements on 16 May took place between 1630 and 1900 UTC at 90 m height along the pattern labelled by F1 to F6 and additionally at 210 m height along the short east-west oriented branches of the flight pattern. The FALCON measurements were accompanied by simultaneous LIDAR measurements of the D-CALM flying at 3 km altitude. Further details on the experiment and the platforms are given in Brüllmer et al. (1991).

To set up the case study which serves as a control run, we have to define the geographical latitude, the boundary conditions, the external parameters which characterize the large-scale field, and to specify the initial conditions for all prognostic variables. The parameters and the boundary-layer profiles have been selected in agreement with the main features observed during the period 1630–1900 UTC on 16th May 1988 in the ARKTIS experiment (Brüllmer et al., 1991). During this period, a northerly wind caused cold air to flow off the ice pack over relatively warm water, where the PBL was quickly modified. Cloud streets formed 40 km or less downstream from the coastline. The cloud layer thickened quite rapidly with downwind distance from the coast and formed a near closed cloud layer at a distance of about 100 km from the ice edge, in which the rolls can be identified as rows of dense dark clouds. Figure 2 shows mean horizontal distributions of air and water temperature $T$ and $T_w$, water vapour mixing ratio $m$, the horizontal wind vector $\mathbf{v}$ and boundary-layer height $h$ derived from the FALCON flights at 90 m height and from LIDAR measurements. Both the air temperature
Fig. 1. ARKTIS 1988 area with flight patterns of the research aircraft FALCON (A–D) on 15 May and FALCON (F1–F6) and DORNIER128 (D1–D8) on 16 May 1988. Open circles and open squares mark the positions of the research vessels POLARSTERN and VALDIVIA on 15 and 16 May, respectively. The thick full line marks the ice edge (after Brümmner et al., 1991).

$T$ and the water vapour mixing ratio $m$ increase downstream at a rate of about 3.5 K per 200 km and 0.4 g/kg per 200 km, respectively. The $T_w$-field exhibits a rather complex structure. However, there is a general tendency for increasing sea surface temperatures towards the southwest. The boundary layer thickens in the downwind direction but additionally shows a crosswind inclination from northwest to southeast. Wind speeds are between 10–18 m/s.

We consider the idealized problem of a geostrophic wind blowing with a crosswind inclination of about 45° with 25 m/s. An east-west oriented infinite straight coastline separates the ice pack from the ocean with surface temperature varying normal to the coast, but uniform parallel to the coast. The model is driven by time-varying, uniform surface boundary conditions and a geostrophic wind parallel to the $y$-axis. In comparing with observations, we have to remember that the computational domain is translating with a fixed speed along the geostrophic wind
Fig. 2. Mean horizontal distributions of air and water temperature T and T', (left panel), water vapour mixing ratio on (middle panel), and wind vector v and boundary layer height h (right panel) derived from 10 km average values (dots) measured at 90 m height by FALCON on 16 May 1988. Isolines of h are based on profile soundings and LIDAR measurements (after Brummet al. 1991).
TABLE I

<table>
<thead>
<tr>
<th>Input parameters for the control run</th>
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<tbody>
<tr>
<td>$f = 1.428 \cdot 10^{-4}$ s$^{-1}$</td>
</tr>
<tr>
<td>$u_s = 0$, $v_s = 25$ m s$^{-1}$</td>
</tr>
<tr>
<td>$\bar{q}(z = z_f, t = 0) = 3.6 \cdot 10^{-3}$</td>
</tr>
<tr>
<td>$\Delta \bar{q}(t = 0) = 1.4 \cdot 10^{-3}$</td>
</tr>
<tr>
<td>$\Gamma_q = 0$ m$^{-1}$</td>
</tr>
<tr>
<td>$\frac{\partial \bar{q}}{\partial t}/z = z_f = \frac{\partial \bar{q}}{\partial T}$</td>
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<tr>
<td>$\frac{\partial \bar{q}}{\partial t}<em>{LS} = \frac{\partial \bar{q}}{\partial t}</em>{LS} = \frac{\partial \bar{q}}{\partial t}<em>{LS} = \frac{\partial \bar{q}}{\partial t}</em>{LS} = 0$</td>
</tr>
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Thus, with a geostrophic wind speed of 25 m/s and an angle of 45° between the coastline and the geostrophic wind direction, 1 hour of integration time corresponds to a traveling distance of about 64 km normal to the coastline. The latitude has been taken to be 79°N. At the shoreline, the lower boundary condition corresponds to an oceanic surface at 272.16 K at saturation ($\bar{q} = 3.6 \cdot 10^{-3}$), where the roughness lengths are equal to 10$^{-3}$ m. The temporal temperature variation of the ocean surface has been chosen equal to $1.875 \cdot 10^{-4}$ K s$^{-1}$, which corresponds to a spatial variation at a rate of $1.06 \cdot 10^{-2}$ K km$^{-1}$. The initial conditions, cloudless and horizontally homogeneous, have been prescribed according to the procedure described in Section 2.3. Except for a superadiabatic surface layer, the initial atmosphere was stably stratified with a lapse rate of $\Gamma_{\text{at}} = 8.88 \cdot 10^{-3}$ K m$^{-1}$. A $\Delta \bar{T}_l = 5$ K difference has been chosen between the temperature at the ocean surface and the first model grid point above the surface at $z = z_1 = 25$ m. The initial moisture profile shows a strong drop in specific humidity from saturation ($\bar{q} = 3.6 \cdot 10^{-3}$) at the surface to a constant value of $2.2 \cdot 10^{-3}$ above (i.e. $\Delta q = 1.4 \cdot 10^{-3}$ and $\Gamma_q = 0$). The initial wind profile is provided by a one-dimensional version of the model. The values of the input parameters for the case study (control run) are summarized in Table I.

The large-scale corrective terms describing the effect of the inhomogeneity of the synoptic-scale meteorological fields have been set to zero, because no attempt has been made to allow the calculated mean profiles to be maintained to the observed ones. Such an adjustment would require specification of the synoptic-scale horizontal and vertical advection with high accuracy. Since we are only interested in understanding the basic features of boundary-layer modification during a cold air outbreak, these complicating factors are excluded. We think, however, that our case study, despite its simplifications, can give valuable insight
into typical properties of the dynamics of boundary-layer rolls during cold air outbreaks which would also apply to a more realistic boundary layer.

Secondly, five sensitivity simulations were performed to clarify the influence of various physical processes and external parameters on the development of coherent structures within the convective boundary layer. Our principal objectives are to determine the respective roles of a mean subsidence, a time-dependent surface heating, a large-scale moisture field, cloud top radiative cooling and latent heat release on boundary-layer modification by considering simulations with and without these effects. The sensitivity runs (1) to (3) were different in that one of the external parameters given in Table 1 is varied while the others are fixed. In this way, we investigated the influence of large-scale mean subsidence (Run 1 with $w_{LS} = -10^{-2} \text{ ms}^{-1}$ at $z = z_{TOP} = 1600 \text{ m}$), a time-dependent surface heating (Run 2 with $(d\Phi/dt)_{z=z_T} = 0$), and a large scale moisture field (Run 3 with $\Gamma_q = -1.074 \cdot 10^{-6}$). Run (4) and (5) use the same external parameters as the control Run. Run (4) has no cloud top cooling whilst Run (5) artificially suppresses latent heat release. In Run 1 a linear profile of mean subsidence has been adopted, corresponding to a large-scale velocity divergence of $6.25 \cdot 10^{-6} \text{ s}^{-1}$, independent of height. The decrease of the total moisture content with height in Run 3 corresponds to a drop in the relative humidity from about 90% at $z = 25 \text{ m}$ to about 20% at $z = 1600 \text{ m}$.

The integrations run over 3600 time steps up to $t = 10800 \text{ s}$. In the following, results are presented for the first two hours of the integration period. The later results differ from those at $t = 7200 \text{ s}$ because the boundary layer and the cloud layer keep growing but do not exhibit significantly different structures.

3.2. CONTROL RUN

3.2.1. Secondary flow structure

The flow fields provide an important indication of the nature of the resolved scale eddies. Figures 3–4 and 6–7 illustrate typical realizations of the secondary flow pattern in vertical and horizontal plan views at times $t = 1800 \text{ s}$ (corresponding to a distance from the coast line of about $y_d = 32 \text{ km}$), $t = 4500 \text{ s}$ ($y_d = 80 \text{ km}$) and $t = 7200 \text{ s}$ ($y_d = 128 \text{ km}$), respectively. The Figures show isopleths of the various fields, where, except for $\tilde{q}_f$ and $\tilde{e}$, horizontal averages have been removed. In addition, the quantities are normalized by using the maxima of the quantities occurring in the cross-section. Solid and dashed lines represent, respectively, positive and negative deviations from the mean; the contour interval is 0.2. Figure 3 shows contours of the vertical component of the velocity field in vertical $x$-$z$ cross-sections located at $y = 3200 \text{ m}$ (left panel) and in horizontal $x$-$y$ sections in the middle of the cloud layer (right panel) at times (a) $t = 1800 \text{ s}$ (at the 350 m level), (b) $t = 4500 \text{ s}$ (at the 400 m level), and (c) $t = 7200 \text{ s}$ (at the 500 m level). Figures 4 and 6–7 represent vertical and horizontal cross-sections for the variables $\tilde{q}_f$, $\tilde{\Theta}_c$ and $\tilde{e}$ for the same levels and times as Figure 3.
Fig. 3. Contour plots of the vertical velocity in vertical x-z cross-sections located at y = 3200 m (left panel) and in horizontal x-y planes (right panel) at times (a) \( t = 1800 \) s (\( z = 350 \) m), (b) \( t = 4500 \) s (\( z = 400 \) m) and (c) \( t = 7200 \) s (\( z = 500 \) m) for the control run. Vertical velocity is normalized by using the maximum of this quantity occurring in the cross-section (maximum indicated). Solid and dashed lines represent, respectively, positive and negative perturbations from the mean; the contour interval is 0.2.

At time zero, the model starts from the equilibrium mean fields already described, except for a small random disturbance applied to the temperature at the lowest level. Its amplitude increases by instability and the boundary layer is found to grow rapidly as the cold air flows off the ice over the relatively warm water.
Coherent structures were identified in this boundary layer. The three-dimensional results show that after about 15 min, convection starts in the form of horizontal roll vortices. Figure 3a shows fields of the vertical velocity at $t = 1800$ s, a short time after the onset of convection. As seen in the $x$-$y$ plane, the $\bar{w}$-eddies are organized into distinct bands oriented about $50^\circ$ to the left of the direction of the geostrophic wind. In comparing the alignment of the bands with the wind direction, an angle of about $29^\circ$ results (the arrows labeled 'wind' in Figure 3 give the...
direction of the surface wind). The lateral wavelength of the bands of about 1300 m in a layer of depth 500 m gives an aspect ratio of the rolls of about 2.6. It is striking from the horizontal field that no regular two-dimensional motions are generated by the model. Although the optical impression is that of elongated or streak-like updrafts, the flow field exhibits considerable variations along the roll axis on scales comparable to the lateral wavelength. Therefore, the assumption of homogeneity along the streamwise direction of two-dimensional studies seems hardly justified on the basis of our three-dimensional results. In the $x$-$z$ cross-section, parts of a roll-shaped circulation are seen, with alternating regions of up and downdrafts. These structures are not steady in form or highly regular though rolls propagating in the direction normal to the bands are the dominant feature. The maximum amplitude of $w$ occurs at $z = 500$ m, which corresponds roughly to the height of the inversion base. At this height, the largest shear is found in the lateral mean wind component, which leads us to the assumption that these structures are related to shear instability. The initial development in the form of elongated updrafts seems to break up after 1 h of integration time. At $t = 4500$ s, the $w$-eddies are more irregular and there is little evidence of any roll structure in this field. After 2 h of integration time, a reorganization of the vertical velocity field in the form of rolls is evident. The eddy elongation which does occur is close to the direction of the geostrophic wind, as shown in Figure 3c. This orientation coincides with values found by Mason and Sykes (1982) under unstable conditions. Most of the energy in the vertical velocity power spectrum is at the 2.1 km wavelength which corresponds to a roll-aspect ratio (wavelength/boundary-layer height) of about 4 which is typical of rolls developing during cold air outbreaks at a greater distance from the shoreline and corresponds to values reported by Brümmer et al. (1991). The maximum values of the $w$-component are in the range 1–2 m/s, and in agreement with experimental and theoretical results (Hein and Brown, 1988; Brümmer et al., 1991; Raasch, 1990; Sykes et al., 1988, 1990). Note that now two centers of maximum vertical velocity occur. The secondary maximum is located at $z = 700$ m just above the inversion base and is associated with the strong shear in the lateral mean wind component. The primary maximum occurs at $z = 350$ m within the convective boundary layer, indicating that the rolls become increasingly more convective in character with distance from the ice edge.

Figure 4 shows the liquid water field from the run, illustrating the initiation of clouds after about 1800 s, i.e., at a distance of about 32 km from the shoreline. The cloud layer thickens quite rapidly, and forms nearly a closed layer at $t = 7200$ s. The clouds (the word cloud refers to the region where $\tilde{q}_l$ is positive) occur in a layer extending between 250–750 m. The maximum liquid water content reaches 0.4 g/kg at $t = 7200$ s, which seems to be reasonable for stratocumulus clouds. The cloud cover varies between 20% at $t = 1800$ s and 100% at $t = 7200$ s. The liquid water content does not exhibit horizontal homogeneous structure but shows a clear positive correlation with the vertical velocity field, i.e., large values of $\tilde{q}_l$ occur simultaneously with strong ascending motions whilst descending mo-
tions are characterized by small values of $\bar{q}_r$. In addition, penetrating updrafts cause an upwelling of the contours of constant liquid water content so that the cloud tops protrude into the mean capping inversion. Therefore, cloud-top height, as well as cloud-base height are variable, typically over a 100 m height interval, within the context of the horizontal average. Simulated downstream variations of boundary-layer heights and cloudiness over the open water are in rough agreement with the observations of Brümmner et al. (1991) but there seems to be an overprediction of mean cloud top heights. This is illustrated in Figure 5, which shows two examples of the DO-228 LIDAR measurements of cloud top heights along the northern and southern crosswind flight legs.

The fields of virtual potential temperature shown in Figure 6 are seen to be well correlated with the field of the vertical velocity. At $t = 1800$ s the field of the vertical velocity is nearly at every level (except at the lowest two levels) out of phase with the temperature perturbation field, i.e., maximum temperature perturbations coincide with strongest downward motions and vice versa, indicating conversion between roll kinetic energy and potential energy, resulting in a deepening of the boundary layer. At $t = 4500$ s and $t = 7200$ s, the correlation between the $\tilde{w}$- and $\tilde{\Theta}_v$-fields is positive in the sub-cloud-layer and in the lower part of the cloud layer, indicating an upward buoyancy flux, and is negative above. The temperature perturbation $(\tilde{\Theta}_v - \langle \tilde{\Theta}_v \rangle)$ is determined largely by vertical advection of the mean temperature $\langle \tilde{\Theta}_v \rangle$. Hence, large values of $\tilde{\Theta}_v - \langle \tilde{\Theta}_v \rangle$ were confined to the inversion layer, the region of strong gradients in $\tilde{\Theta}_v$. Maximum temperature perturbations are found to be in the range between 0.5–1.5 K. A negative correlation between the fields of virtual potential temperature and liquid water content near cloud top is also apparent. Virtual potential temperature is higher than ambient only in the lower part of the cloud layer and becomes appreciably lower in the upper part where evaporation of cloud droplets predominates.

The boundary-layer vortex roll structure is also portrayed in the field of SGS-turbulent kinetic energy which shows up clearly in Figure 7. High values of SGS-turbulent kinetic energy are found in the updraft regions of the clouds, indicating significant mixing across the cloud boundaries. Throughout the rest of the domain, the SGS-turbulent kinetic energy is confined to the near surface region, indicating the stable nature of the region between the clouds. This picture confirms the results of LeMone (1976). She found that the circulation of horizontal roll vortices in the thermally unstable PBL influences the distribution of turbulence, with turbulence variances and fluxes concentrated in regions of positive roll vertical velocity. This turbulence modulation was found to be the result of the redistribution of turbulence-producing elements by the roll mean circulation.

### 3.2.2. Mean profiles

Mean values over all points in one horizontal plane are denoted by brackets, e.g., $\langle \Psi \rangle = \langle \Psi \rangle(z)$ denotes a vertical mean profile of any quantity $\Psi$. Double primes denote local deviations from these mean values. For example, $\Psi''$ represents
Fig. 5. Horizontal distribution of cloud top heights from LIDAR measurements of D-CALM from about 3 km altitude. Top panel: flight leg F1-F2 in the broken cloud region at about 50 km distance from the ice edge. Bottom panel: flight leg F5-F6 in the region with complete cloud cover at about 250 km distance from the ice edge (after Brümmer et al., 1991).
Fig. 6. As Figure 1, except for the virtual potential temperature anomalies.

\[ \Psi - \langle \Psi \rangle. \] Total variances and fluxes include resolved-scale and subgrid scale contributions, since, for example,

\[ \langle u'^2 \rangle = \langle u'^2 \rangle + \langle u''^2 \rangle \]
\[ \langle w'' \Theta'' \rangle = \langle w'' \Theta'' \rangle + \langle w'' \Theta'' \rangle. \]

To obtain a compromise between statistical and local representation, the profiles
were averaged over 100 time steps (5 min), at \( t = 1800 \) s and over 300 time steps (15 min) at \( t = 4500 \) s and \( t = 7200 \) s, respectively, corresponding to a time interval in which the boundary-layer growth is less than 50 m.

Figures 8a–8f show mean boundary-layer profiles of \( \langle \bar{u} \rangle, \langle \bar{v} \rangle, \langle \bar{\Theta} \rangle, \langle \bar{q} \rangle \) and \( (\partial \bar{\Theta}/\partial t)_R \). Solid lines are used to denote the initial profiles and dashed line patterns to denote the profiles after 1800, 4500 and 7200 s of integration time, respectively. The initial longitudinal wind component \( \langle \bar{v} \rangle \) is characterized by a
Fig. 8. Vertical mean profiles of the lateral velocity component ($\bar{u}$) (a), of the longitudinal velocity component ($\bar{v}$) (b), of the virtual potential temperature ($\bar{\Theta}_v$) (c), of the specific humidity ($\bar{q}_w$) (d), of the liquid water content ($\bar{q}_l$) (e), and of radiative cooling ($\bar{q}_r(\bar{\Theta}_v/\bar{u})$) (f) for the control run. Solid lines are used to denote the initial profiles, dashed line patterns to denote the profiles at $t = 1800$, $4500$ and $7200$ s, respectively.
weak maximum at $z = 400$ m and an inflection point may be noted in the lateral $\langle \hat{u} \rangle$-wind component at a height of about 350 m. The initial profile of virtual potential temperature $\langle \tilde{\Theta}_v \rangle$ is characterized by a temperature difference between the ocean surface and the overlying air and a constant lapse rate above. Likewise, the initial humidity profile $\langle \tilde{q}_v \rangle$ shows a humidity jump in the surface layer and a constant value above. In the course of integration, the profiles of the basic boundary-layer variables demonstrate a nearly well mixed convective cloud-topped boundary layer. The divergence of the longitudinal momentum transport associated with the subgrid-scale eddies and with the rolls reduces the mean vertical gradients of $\langle \hat{v} \rangle$-momentum in the boundary layer but produces a wind jump from the wind in the mixed layer to the geostrophical wind above. With respect to the mean lateral wind component, we note that $\langle \hat{u} \rangle$-momentum is well mixed in the boundary layer and that the inflection point height lifts with time (occurring in the vicinity of the capping inversion) paralleling boundary-layer growth. Surface heating leads to the formation of a mixed layer near the sea surface, surmounted by a conditionally unstable cloud layer. This layer is capped by an inversion whose height grows in time (or with distance from the shoreline). The profile of specific humidity shows a strong drop from saturation at the surface to a nearly constant value in the mixed layer and a linear decrease in the cloud layer. In addition, due to the surface moisture and heat flux, the boundary layer is moistened and warmed reducing both the moisture jump and the temperature difference between the sea surface and the overlying air. These features are in agreement with the observations as can be seen in Figure 2. The mean liquid water content shown in Figure 8e has a peak value of about $1.6 \cdot 10^{-4}$ kg/kg at $t = 7200$ s. The cloud base is located at $z = 250$ m; from this height, $\langle \tilde{q}_v \rangle$ increases adiabatically up to 500 m. In the inversion zone, $\langle \tilde{q}_v \rangle$ is subadiabatic, which is a typical feature owing to the entrainment of dry air from above and the subsequent evaporation of cloud droplets. The cloud-top cooling $\langle \hat{\Theta}_v / \hat{\tau} \rangle_R$ is distributed over a depth of about 250 m, which appears to be a rather deep layer, but seems to be reasonable realizing that $\langle \hat{\Theta}_v / \hat{\tau} \rangle_R$ represents a space-time average and that cloud-top heights exhibit a wave-like undulation with differences in height of up to 100 m. The maximum cooling rate of 2.4 K/h occurs at level $z = 600$ m, corresponding to the level of maximum liquid water content. This value is strongly dependent on the vertical mesh size and the cooling peak is localized near the surface of clouds. The fact that the ocean surface emits radiation at higher temperatures than the cloud is also seen in Figure 8f. It leads to a slight heating at the grid points shortly above the cloud base.

Important variables for describing the turbulence structure of any boundary layer are the second-order moments. Vertical profiles of resolved-scale horizontal and vertical velocity intensities as well as the variance of the resolved-scale virtual potential temperature fluctuations are shown in Figure 9a–9d at times $t = 1800$ s (solid), $t = 4500$ s (short-dashed) and at $t = 7200$ s (long-dashed). At $t = 1800$ and 4500 s, the profiles of $\tilde{u}^\prime$ and $\tilde{w}^\prime$-variance exhibit maxima in the upper part of the
boundary layer, not far from the inflection point in the mean lateral wind component, where the largest shear is present. These profiles are typical for situations in which shear-generated vortices due to inflection-point instability are met (e.g., Chlond, 1987). The longitudinal velocity variance is seen to maintain a nearly constant value in the mixed layer but shows a maximum near the boundary-layer top, which can partly be explained by the fact that the profile of $\tilde{v}''$-variance is produced by the adjustment from $\tilde{v}$ to $v_g$ in the thin inversion layer. At $t = 7200$ s, an absolute maximum of $\tilde{w}''$-variance occurs in the middle of the boundary layer. This feature agrees with the idea that the rolls become increasingly more convective in character with distance from the shoreline. A relative maximum of $\langle \tilde{w}''^2 \rangle$ again occurs near cloud top indicating mechanical energy production. The variance of the horizontal velocity fluctuations shows two maxima in the boundary layer. The first maximum occurs near the surface where the wind shear is important,
and the second maximum appears in the entrainment zone owing to the inflection-point instability. In contrast to previous two-dimensional roll theories (e.g., that of Mason and Sykes (1982) or that of Chlond (1987)) but in accordance with the observations, the model predicts a resolved-scale variance ratio \( \langle \theta'^2 \rangle / \langle u'^2 \rangle \) of the order one, while two-dimensional models predict a ratio of \( \langle \theta'^2 \rangle_{\text{max}} \) to \( \langle u'^2 \rangle_{\text{max}} \) between 4:1 and 10:1. This supports the suggestion that three-dimensional effects reduce the resolved-scale variance ratio \( \langle \theta'^2 \rangle / \langle u'^2 \rangle \) because, as a result of the pressure-velocity-gradient correlation \( \langle \bar{p}'' \partial \bar{v}' / \partial y \rangle \), coupling of \( v'' \)-momentum to \( u'' \)-, and \( w'' \)-momentum is stronger than two-dimensional models suggest. The effect of this correlation is to produce an equipartition of resolved-scale energy since the transfer of energy, via pressure forces, is from the high-intensity to the lower intensity component.

The variance of the resolved-scale virtual potential temperature fluctuations is shown in Figure 9d. Temperature fluctuations are produced by the negative product of heat flux and temperature gradient. This product is large at the inversion but is small in the middle of the boundary layer. This explains the general shape of the profile, as in the mixed layer, no large fluctuations are possible for air parcels displaced in the vertical direction. On the other hand, the \( \Theta'' \),-variance reaches a maximum at the inversion. This maximum is associated with the development of increased stratification at the inversion because in the entrainment zone, even small vertical motions cause large fluctuations because of the strong vertical gradients there. This tendency has been confirmed from experimental and numerical studies for dry and moist convective boundary layers (e.g., Brümmer, 1985; Brümmer et al., 1991; Mason, 1989; and Sykes et al., 1990).

Vertical profiles of longitudinal and lateral momentum flux are shown in Figures 10a–10b. These profiles, as well as those given in the following figures, represent time averages and, except when indicated, the flux curves include the resolved-scale values and their subgrid-scale counterpart. The calculated longitudinal momentum flux \( \langle \bar{v}' \bar{w}' \rangle \) has its maximum close to the sea surface and decreases rapidly with height to values near zero at the top of the boundary layer. At \( t = 1800 \) and \( 4500 \) s, the total downward longitudinal momentum flux is nearly entirely transported by the parameterized SGS-eddies whilst at \( t = 7200 \) s, about 40% of the total longitudinal momentum flux existed on the subgrid-scale at \( z = 200 \) m; this fraction dropped to about 20% for \( z > 300 \) m. The lateral momentum flux is generally smaller than the longitudinal one and the major contribution comes from the subgrid-scale. The flux decreases with increasing distance from the sea surface and shows a minimum in the upper half of the boundary layer. In the lower part of the boundary layer, the divergence of this flux tends to increase \( u \)-momentum and vice versa in the upper part. Therefore, the action of the lateral momentum flux divergence is to diminish the wind difference above and below the inflection point to relieve the vorticity maximum.

Figure 10c shows the profile of virtual potential temperature heat flux (buoyancy flux) together with the SGS-contribution. The buoyancy transport is nearly totally
due to SGS mixing during the first 1.5 h of integration except in the inversion zone where the resolved-scale fraction appears dominant. Afterwards, a considerable part of the transport of virtual potential temperature was taken over by the resolved-scale eddies. At \( t = 7200 \text{s} \), the total buoyancy flux exhibits a typical structure. It is close to linear up to the cloud base and is caused by surface heating owing to the temperature difference between the sea surface and the overlying air of about \( \Delta \Theta \approx +2.9 \text{K} \). The linear profile implies a constant heating rate in the mixed layer and indicates stationarity of the turbulent state in the mixed layer. Inside the cloud, the virtual potential temperature heat flux increases to values of \( \approx 50 \text{W m}^{-2} \) and attains a maximum in the middle of the cloud layer. Above, the heat flux again decreases slowly throughout the upper cloud layer, changes sign at \( z = 600 \text{m} \) and becomes slightly negative in the inversion zone because
warmer and drier air from above is transported against thermal stability into the boundary layer (i.e., due to entrainment). The maximum negative entrainment heat flux amounts to $\approx -10 \text{ W m}^{-2}$. The dashed curves represent the SGS contributions. At the surface, the resolved-scale vertical velocity vanishes and therefore, all the flux is transported by the SGS contribution at this level. Above this level, the mean SGS flux is very small and becomes negative near the inversion. Our results show that there is a very substantial positive buoyancy flux in the upper part of the boundary layer. The reason for this is twofold. First, radiative cooling, which is most efficient at cloud top, leads to a positive buoyancy flux there for the same reason that heating at the bottom of the boundary layer leads to a positive buoyancy flux there, i.e., to compensate the loss of heat due to radiative cooling, a large positive heat flux is established inside the cloud. In addition, conversion of water vapour to liquid water releases latent heat and therefore, creates a positive buoyancy flux.

The water vapour flux (Figure 10d) was mainly due to boundary-layer turbulence for the first phase of the cold air outbreak, before a substantial part of the water vapour transport was taken over by the resolved-scale roll circulation. The profile of moisture flux derived from the simulation is positive (i.e., directed upward) over the total area. Evaporation at the ocean surface increases the moisture content in the surface layer, which is then mixed by turbulence and the roll vortices to upper regions. Therefore, the flux has its maximum close to the sea surface and decreases with height to small positive values at the top of the boundary layer where boundary-layer moisture is diluted by entrainment of dry environmental air. However, as this dilution is smaller than the import from the surface, the total amount of moisture increases in the whole boundary layer during the course of integration. At $t = 7200$ s the parameterized fraction of the moisture flux appears negligible except in the lower 150 m and at the top of the boundary layer.

Production terms in the budget equation for resolved-scale turbulence energy $\langle E_R \rangle = \frac{1}{2} \left( \bar{u}^2 + \bar{v}^2 + \bar{w}^2 \right)$ are shown in Figure 11a and 11b as function of height at 1800, 4500 and 7200 s. Positive values contribute towards growth of $\langle E_R \rangle$ and negative values contribute towards decay. The production of roll kinetic energy by buoyancy $g/T_0 (\bar{w}' \bar{\Theta}''')$ describes the flow of energy from the reservoir of potential energy to roll kinetic energy by means of a positive heat flux whilst the shear production term $\langle \bar{w}' \bar{w}'' \rangle \frac{d\langle \bar{w}' \rangle}{dz} + \langle \bar{v}' \bar{w}'' \rangle \frac{d\langle \bar{v}' \rangle}{dz} \frac{d\langle \bar{v}' \rangle}{dz}$ arises from the interaction of the Reynolds stress and the mean wind components. At $t = 1800$ s, the strength of the shear term is apparent, especially at the 300 m height. This is not surprising considering the large wind shear in the mean lateral wind component at this height. Thus, we conclude that during the first phase of the cold air outbreak, the rolls are produced by dynamic (inflection point) instability, which corresponds to the dynamic energy source for rolls as described by Brown (1970). As seen in Figure 11b, a great portion of this energy gain is converted into potential energy by buoyant destruction of energy since the rolls have to perform work against stable stratification (i.e., a great portion of shear-generated turbulence is used to lift the
Fig. 11. Average profiles of the energy production of resolved-scale turbulence energy \( \langle E_r \rangle = \frac{1}{2} \langle \dot{u}^2 + \dot{v}^2 + \dot{w}^2 \rangle \) by wind shear of the basic flow \(- \langle \dot{v} \dot{w} \rangle \frac{d\langle \dot{u} \rangle}{dz} - \langle \dot{u} \dot{w} \rangle \frac{d\langle \dot{u} \rangle}{dz} \) (a) and by buoyancy \( g/\theta_0 \langle \dot{v} \dot{\theta} \rangle \) (b) for the control case. Other features as in Figure 6.
boundary layer by entrainment of warm stable environmental air from above). At $t = 4500$ s, the shear stress as well as the buoyancy production terms are seen to be rather small, indicating a transition stage. At a greater distance from the shoreline (i.e., at $t = 7200$ s), there exists a broad maximum in the buoyancy production in the sub-cloud and cloud-layer, suggesting that buoyancy is now the main driving force for the rolls. The shear production term is seen to have been rather unimportant in the middle part of the boundary layer, but turns out to be dominant in the resolved-scale turbulent energy equation near the ground and near the inversion. Our results show a good qualitative correspondence with measured values of the kinetic energy production terms which are based on FALCON profile soundings on 16 May 1988. As can be seen in Figure 12, the observations also support the idea that the secondary flow structures are formed by dynamic processes during the first phase of the cold air outbreak and that the thermal convection becomes dominant later on. However, there are some discrepancies between the observations and the simulated results. In particular, the model tends to overestimate the magnitude of the energy generation terms. This apparent discrepancy may at least be partly traced back to the fact that the measured fluxes only include contributions from motions with horizontal scales up to 1 km whilst the model results include contributions from motions with horizontal scales up to 6 km. In summary, inspection of the generation terms in the roll

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**Fig. 12.** Profiles of kinetic energy production by wind shear of the basic flow $-u'w'$ and by buoyancy $g/\Theta w'\Theta'$ derived from 1 km long horizontal subsections of the FALCON profile flights on 16 May 1988. Profiles in the left panel represent averages over the profiles at F1 and F2 in the region with broken clouds and the profiles in the right panel are averages over the profiles at F3 to F6 in the region with a closed cloud deck. The averages are performed relative to cloud base $z_B$ and cloud top $z_T$ to maintain characteristic structures (after Brümmer et al., 1991).
kinetic energy budget equation permits one to quantify the processes leading to
the formation of rolls. We found that during the first phase of the cold air outbreak
the rolls were produced by dynamic instability in the presence of strong vertical
wind shear but became increasingly more convective in character with distance
from the shoreline owing to convective instability supplying energy at the roll
wavelength. However, the large value of the shear stress term at the top of the
boundary layer leads us to conclude that dynamic instability still plays an impor-
tant role in determining the roll structure during the later stage.

3.3. Sensitivity runs

In order to examine the sensitivity of the simulated roll vortices to various physical
processes, sensitivity tests were performed. The control case was re-run with

(a) large-scale mean subsidence (Run 1 with \( w_{LS} = -10^{-2} \text{ ms}^{-1} \) at \( z = z_{TOP} = 1600 \text{ m} \))
(b) no time-dependent surface heating (Run 2 with \( (\partial \tilde{\theta}_h / \partial t)_{z = z_b} = 0 \))
(c) a modified large-scale moisture field (Run 3 with \( \Gamma_q = -1.079 \cdot 10^{-6} \text{ m}^{-1} \)
corresponding to a linear decrease of the total water content \( \tilde{q} \) with height
at time \( t = 0 \text{ s} \) from \( \tilde{q} = 2.2 \cdot 10^{-3} \text{ kg/kg} \) at \( z = 25 \text{ m} \) to \( \tilde{q} = 0.5 \cdot 10^{-3} \text{ kg/kg} \)
at \( z = 1600 \text{ m} \))
(d) no long-wave radiative effects (Run 4) and
(e) latent heating due to condensation and evaporation artificially turned off
(Run 5).

Figure 13 shows vertical profiles of resolved-scale vertical velocity variance (a),
resolved-scale virtual potential temperature heat flux (b), and of mean liquid water
content (c) for the different runs at time \( t = 7200 \text{ s} \) during the convective regime.
This time period has been chosen because the boundary-layer profiles for the
mean quantities (apart from changes in boundary-layer temperatures and moisture
contents in Runs 2 and 3 arising from different initial and boundary conditions)
and for the second-order moments appear very similar for the various cases during
the first 1.6 h of integration time.

As seen in Figure 13a, the omission of surface heating, latent heat release,
 radiative cooling, large-scale subsidence or less water vapour produce less vigorous
boundary-layer eddies. The greatest reductions in the vertical velocity intensity
occur in Runs 3 and 5 where \( \langle \tilde{w}'' \rangle_{\text{max}} \) has dropped from 0.3 to 0.12 and 0.13 m² s⁻²,
respectively. The physical reason for this effect is the lowered cloud activity (Run
3) and the absence of the energy source (Run 5) due to latent heat release,
respectively. In Run 3 owing to the initial decrease of the total water content with
height, entrainment of relatively drier air from the region just above the inversion
tends to dry out the boundary layer resulting in a conversion rate from water
vapour to liquid water (and hence a heat release) which is much less than in the
reference run. Therefore, condensation provides a significant energy source caus-
ing more intense boundary-layer eddies. This feature agrees with the idea that
ascending buoyant plumes receive an additional vertical impulse due to heat release. Moreover, Figure 13a depicts that the lack of significant cloud top cooling (Run 4), the absence of time-dependent surface heating (Run 2) as well as large-scale subsidence also severely weakened the system as a whole. The comparison
reveals a drastic decrease of vertical velocity intensity $\langle \bar{w}^2 \rangle_{\text{max}}$ (0.14 m$^2$ s$^{-2}$ (Run 4), 0.15 m$^2$ s$^{-2}$ (Run 2) and 0.22 m$^2$ s$^{-2}$ (Run 1) versus 0.30 m$^2$ s$^{-2}$ in the control run).

The (de)-stabilizing effect of the various physical processes investigated is also reflected in the profiles of virtual potential temperature heat fluxes (Figure 13b). In the case of large-scale subsidence (Run 1) and in the absence of time dependent surface heating (Run 2), the buoyancy flux shows a uniform decrease of 10 and 40% at all levels, respectively. The great reduction of the virtual potential temperature heat flux in the case of a spatially-homogeneous ocean surface temperature is reasonable, realizing that due to the input of heat at the ocean surface the boundary layer is warmed with downwind distance from the shoreline, thereby diminishing the air-sea temperature difference. In the control run, the latitudinal temperature variation of the ocean surface partly counterbalances boundary-layer warming. As a consequence, a much higher surface heat flux is obtained in the control run (100 versus 60 W m$^{-2}$) than in Run 2 leading to more unstable conditions in the surface layer. The surface-layer instability is therefore seen as a feature which enhances the ability of the cloud circulation to 'pull' air upwards and to influence the overall potential energy of the atmosphere due to enhanced vertical convective transport. The reduced surface instability in Run 2 cuts off part of the internal energy input into the model atmosphere, and thus a less vigorous roll circulation is found after 2 h of integration time. Smaller values of the buoyancy fluxes are also found in the sensitivity Runs 3 and 6, especially in the upper
part of the boundary layer because as explained in the preceding paragraph, the convectively generated updrafts do not provide a corresponding latent heat release. Radiation effects have not been included in many studies of the convective boundary layer (e.g., Sykes et al., 1988; Raasch, 1990). The underlying assumption was that the surface fluxes should dominate boundary-layer development, justifying the neglect of radiation. However, our numerical simulation of boundary-layer vortices (Run 4) has revealed a surprising sensitivity to longwave radiative cooling/heating. We find that longwave radiation from the cloud top can produce strong cooling and hence significantly influence the dynamics of clouds. The effect of radiative cooling is twofold. First, the long-wave radiative flux divergence existing in the entrainment zone serves to cool the cloudy air in that zone and therefore, promotes entrainment by reshaping the temperature profile there (Kahn and Businger, 1979, refer to this process as ’direct’ entrainment). Second, that part of the radiative flux divergence concentrated below the entrainment zone (i.e., within the upper cloud layer inside the convective boundary layer) will, on the other hand, produce kinetic energy just as does heating at the ground and will, therefore, be one of the driving mechanisms of boundary-layer turbulence. Our calculation reveals that the latter process is of primary importance in the model atmosphere. We find that the increase of resolved-scale vertical variance in the control run (including radiation) is linked to an increase in the vertical flux of virtual potential temperature, which is one of the main sources of kinetic energy. Compared to Run 4, this quantity is at time $t = 7200$ s approximately doubled in the control run inside the cloud layer (see Figure 13b). Therefore, we conclude that in the case of the convective boundary layer during cold air outbreaks, the cloud-radiation interaction increases cloud activity and related turbulent quantities due to destabilization of the cloud layer.

The impact of the absence of cloud-top cooling is also documented in Figure 13c showing profiles of the liquid water content. The lack of radiative effects has severely weakened the system as a whole. The peak liquid water content has dropped from 0.16 to 0.10 g/kg and the fractional cloud cover is smaller (0.80 versus 1.00) than in the reference case. In comparing liquid water contents and fractional cloud cover from Runs 1, 2 and 5 with those of the control run, we find that the large-scale subsidence, the absence of time dependent surface heating and the lack of latent heating have only a small influence upon cloud structure. For example, as a consequence of warming the convective boundary layer, large-scale subsidence leads only to a weak reduction in the liquid water content (0.14 versus 0.16 g/kg) and in the fractional cloud cover (0.97 versus 1.0). In contrast, a large-scale vertical moisture gradient (Run 3) produces dramatic changes in liquid water and the whole boundary-layer structure. In this case, the entrainment of warmer and drier air from aloft prevents the formation of a complete cloud cover. Therefore, the gradient in the $\tilde{q}$-profile results in a drying of the boundary layer, which is documented by lifting of the cloud base (350 versus 250 m) relative to that seen in the control run. In addition, the resulting fractional cloud cover is far smaller.
RUN 3

VERTICAL VELOCITY

TIME = 2 h 0 min 0 s

MAX = 0.1035E+01
MIN = -0.8431E+00
LEVEL = 500 m

LIQUID WATER

TIME = 2 h 0 min 0 s

MAX = 0.2324E-03
MIN = 0.0000E+00
LEVEL = 500 m

Fig. 14. Realizations of the vertical velocity (top panel) and the liquid water content fields (bottom panel) in vertical (at y = 3200 m) and horizontal plan views (at z = 500 m) for Run 3 at t = 7200 s.

than in the reference case (0.32 versus 1.00) and the peak area-averaged liquid water content has dropped from 0.16 to 0.018 g/kg.

Figure 14 shows realizations of the vertical velocity and the liquid water content fields for Run 3 in horizontal (at z = 500 m) and vertical (at y = 3200 m) plan views at time t = 7200 s. The flow field consists of much more regular and steady eddies than those of the control run. The flow is again dominated by boundary-layer rolls with a wavelength of about 1070 m. The updrafts have about the same width as the downdrafts. The maximum concentrations of the liquid water content are located near the top of the strongest updrafts and move steadily from right to left in the illustrations shown in Figure 14. There is a tendency for clouds to trail behind the updrafts. The lower relative humidity in the free atmosphere (owing to the prescribed initial decrease of the total water content in Run 3) results in
reduced cloudiness in the boundary layer relative to that seen in the control run. Thus, the basic mechanism sustaining the cloud-free areas between the updrafts is the mixing of saturated moist cloud air with overlying warmer and drier air. This indicates that under certain environmental conditions, an individual cloud is capable of drying out a large area of the surrounding layer through cloud-top entrainment and thus is able to produce significant gaps between the clouds. Therefore, these results suggest that the large-scale vertical moisture gradient indeed has a strong influence on cloud morphology as well as on the turbulence structure of the convective boundary layer during cold air outbreaks.

4. Conclusions

Results of three-dimensional numerical calculations of boundary-layer roll vortices for a variety of conditions have been presented, using the Large-Eddy-Simulation (LES) technique. The main advance over previous modelling efforts has been to take into account three-dimensional circulations and to include most of the physical processes occurring in a moist boundary layer. Apart from the aspects of the turbulence closure, three main physical processes have been taken into account: a water cycle (including a subgrid scale condensation scheme), infrared radiative cooling in cloudy conditions and the perturbations of the local fields by large-scale subsidence.

As a principal conclusion, it can be stated that the three-dimensional simulation of the boundary-layer eddy structure under conditions approximating those of the cold air outbreak of a particularly interesting case of ARKTIS 1988 (16 May 1988) exhibits cloud structure similar to that observed. Though in some respects the model simulations differ from the ARKTIS actual meteorological situations, typical characteristics of the dynamics of boundary-layer flow during cold air outbreaks are revealed by the model. These include: (1) the boundary layer grows rapidly as the cold air flows off the ice over the relatively warm water; (2) organized structures in the form of quasi-two-dimensional roll vortices are identified in this boundary layer; (3) apart from the entrainment zone, vertical transports of heat and moisture are due to subgrid-scale turbulence for the first phase of the cold air outbreak, before the rolls cause substantial mixing of these quantities as the boundary layer deepens; (4) cloud streets with a wavelength of about 2 km arc formed 40 km or less down-stream from the shoreline. The cloud layer thickens quite rapidly with downwind distance from the coast and forms a nearly continuous cloud layer at a distance of about 100 km from the ice edge, in which the rolls can be identified as rows of denser clouds; (5) roll vortices that form at the beginning have different characteristics than those that appear later on.

The discussion on the mechanisms by which the rolls obtain energy has clarified the relative importance of the various physical processes in the production and maintenance of rolls. We found that during the first phase of the cold air outbreak, the rolls forming within the boundary layer are mainly driven by a dynamic
instability in the presence of a strong vertical wind shear but become increasingly more convective in character with distance from the shoreline owing to the thermal instability supplying energy at the roll wavelength.

Attention has been given to the dependence of the results upon various physical processes and factors influencing the simulations: large-scale subsidence, time-dependent surface heating, a large-scale vertical moisture gradient, radiation and latent heat release. Based on the sensitivity simulations, the following conclusions can be drawn:

- Realistic values of large-scale subsidence are too small to allow more than a small influence on the cloud-topped convective boundary layer. Large-scale subsidence results in a slight decrease in cloud liquid water contents and resolved-scale energy levels.
- As a consequence of a spatially homogeneous ocean temperature, a lowered surface-layer instability reduces the kinematic surface heat flux and thus a less vigorous roll circulation is found after two hours of integration time.
- The presence of a large-scale vertical moisture gradient prevents the formation of a complete cloud cover. In this case, clouds form only in the updrafts of the dominant rolls. The key ingredient for such an organization appears to be the decrease in equivalent potential temperature at the top of the boundary layer in conjunction with the ability of clouds to entrain warm, dry air from aloft in order to give significant gaps between the cloud streets. Therefore, our calculations give grounds for believing that the structure of the initial moisture field indeed has a strong influence on the cloud morphology and ultimately controls the fractional cloud cover.
- The major feature of radiative forcing is seen to be strong cloud-top cooling. This leads to enhanced destabilization of the upper cloud layer, which in turn results in increased cloud activity and related turbulent quantities.
- Latent heating due to condensation provides a significant energy source causing more intense boundary-layer eddies.

Although the results refer to only one case of boundary-layer modification in a cold air outbreak, we think that our simulation may serve as an instructive example giving insight into different interacting physical processes. In addition, the sensitivity simulations undertaken as a test of the response of the model to a change in a significant physical process or parameter indicate that the model indeed responds in a physically plausible manner. Such sensitivity tests are in fact one of the principal objectives of numerical models, and make those a powerful tool to provide insight into the underlying dynamics.

A major criticism of the present simulations is that the model utilizes a rather crude vertical resolution. This drawback may be overcome in the future by increasing the resolution, especially within the capping inversion using adaptive grid techniques. A vertical grid increment of 10 instead of 50 m in this region would (a) reduce the truncation error, (b) allow more accurate estimates of entrainment
fluxes, (c) ensure a better representation of the wind shear, the temperature and moisture jump at the top of the boundary layer and, (d) allow a more accurate treatment of radiative processes occurring in the top layers of clouds.

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References


