The Interdiurnal Variation of Summer Cumulus Convection over the Kanto Plain in Japan

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Abstract

Cumulus convection occurs over the Kanto plain during the afternoons of fair summer days. By investigating cloud amount data, it was shown that the afternoon cloud amount of a given day and that of the following day are not independent of each other. If the cloud amount is small on a given day, the cloud amount tends to be greater on the following day, and vice versa. It can be concluded that fair weather does not tend to persist from day to day, when the summer atmosphere is stable.

The interdiurnal variation was simulated in two-dimensional numerical experiments. Days with strong and weak convection appeared quasi-periodically, even though the complete diurnal boundary conditions were given corresponding to the sea and land breezes. By examining the results of the numerical experiments, it was shown that the interdiurnal variation resulted from the differences in the vertical profiles of potential temperature and water vapor mixing ratio in the lower troposphere.

Furthermore, these differences in the vertical profiles were compared with those in aerological data, in order to examine the above results of the interdiurnal variation.
1 Introduction

Over the tropical ocean, especially in the western Pacific, two-day period oscillations of cumulus convection have been observed (cf. Takayabu, 1994a,b). According to Chen and Houze (1997), the two-day period oscillations are caused by surface-cloud-radiation interactions. When the intraseasonal oscillations (cf. Madden and Julian, 1972) are in a convectively active phase, a large convective system having a nearly 1-day life cycle may develop (Nakazawa, 1988). Such a large convective system fills the surface boundary layer with air having a lower moist static energy due to the down draft. The sea surface is largely shaded from sunshine by the resulting clouds. In addition to the changes in the sea surface conditions, short-wave absorption at the cloud top heats the upper troposphere and stabilizes the free atmosphere. Therefore, on the following day, no large convective system tends to develop in the same area, but rather occurs in a region where no large convective system was present during the previous day. Takayabu (1994a,b) demonstrated that components having periods of 1.5 to 2.5 days are dominant in the equatorial cloud activity of the western Pacific during the December-January-February season, and that such cloud systems have characteristics of an equatorial meridional mode n=1 westward-propagating inertia-gravity wave.

In the present paper, we focused on the interdiurnal variations of cumulus convection, occurring over the Kanto plain in summer under fair weather conditions. As will be mentioned below, the quasi two-day period oscillations appear not only in tropical regions but also in the midlatitudes. In section 2, cloud amount data over the Kanto plain in Japan are analyzed in order to detect interdiurnal variations of summer cumulus convection. In section 3, interdiurnal variations are simulated by simple two-dimensional numerical experiments of the sea and land breezes, to examine the causes of the two-day period oscillation. In section 4, results of the numerical experiments are compared with aerological data over the Kanto plain. Conclusions are given in section 5.

2 Analysis of cloud amount data

Cloud amount data observed by Japan Meteorological Agency were used to detect interdiurnal variations of cumulus convection.

In the following, the values of cloud amount were given in tenths of sky cover. The years 1962, 1965, 1975, 1978, 1984, 1985, 1990, 1994 and 1995 were chosen in order to analyze the cloud amount data. In these selected years, the mean value of total sunshine duration at Tokyo, Kumagaya, and Maebashi during August was greater than 220 hours. Therefore, the weather during these periods was considered fair. Use is made of cloud amount at Tokyo, Kumagaya, and Maebashi during the period July 16 - September 15 of the summer of these years. Figure 1 shows the locations of the three observational sites. These sites
were selected as representative of the region affected by the sea breeze from the Tokyo Bay. Observations were performed every 3 hours at Tokyo and Maebashi, and at 0900, 1500, 2100 LST at Kumagaya. After averaging the values of cloud amount at the three sites, the time-mean values for the morning (0300 - 0900 LST) and for the afternoon (1200 - 2100 LST) of each day were calculated. In figure 2, the values of cloud amount for the afternoon of the first day are plotted along the abscissa versus those on the second day along the ordinate. Only the values for cases in which the mean values of cloud amount are less than 4 during the 3 consecutive mornings have been taken into account. In all of these cases, there was no precipitation during the morning. Furthermore, in most of the cases, the Kanto plain was strongly covered by the subtropical anticyclone. Note here that the total number of cases is 22. The condition of three consecutive days with morning having cloud amount of less than 4 is rather restrictive.

Table 1 is a contingency table for the results shown in Fig. 2. Here, $C_1$ and $C_2$ represent the afternoon cloud amounts on the first day and the second day, respectively. If the cloud amount during the afternoon of the first day is less than 5, the cloud amount on the second day is greater than 5 for eight out of 17 cases. On the other hand, if the afternoon cloud amount is greater than 5 on the first day, that on the second day is less than 5 for all of the 5 remaining cases. We can remark that the former ratio is higher than the latter when the cloud amount on the second day is great.

Hypothesis testings were performed to examine the significance of the tendency mentioned above. First, the null hypothesis $H_0$ and the alternative hypothesis $H_1$ were set as follows.

\[ H_0 : p_1 = p_2, \]  

and

\[ H_1 : p_1 > p_2, \]

where $p_1$ is the possibility of $C_2 \geq 5$ under the condition of $C_1 < 5$ and $p_2$ the possibility under the condition of $C_1 \geq 5$, respectively. Both can be estimated as

\[ \hat{p}_1 = \frac{y_{12}}{n_1}, \]

and

\[ \hat{p}_2 = \frac{y_{22}}{n_2}. \]

Here, $n_1$ and $n_2$ are the numbers of cases in which $C_1 < 5$ and $C_1 \geq 5$, respectively. $y_{12}$ and $y_{22}$ are the numbers of cases in which $C_2 \geq 5$ under the conditions of $C_1 < 5$ and $C_1 \geq 5$, respectively. In this case, $n_1$ and $n_2$ were assigned to
\[ n_1 = 17, \]
\[ n_2 = 5. \]

Under the condition of \( H_0 \), the expected value and the variance of \( p_1 - p_2 \) are
\[ E(\hat{p}_1 - \hat{p}_2) = 0, \]
and
\[ V(\hat{p}_1 - \hat{p}_2) = V(\hat{p}_1) + V(\hat{p}_2) = \left( \frac{1}{n_1} + \frac{1}{n_2} \right) \times p(1 - p), \]
where \( p \) is the common possibility of \( C_2 \geq 5 \) and is written as
\[ p = p_1 = p_2. \]

Using the central limit theorem, the distribution can be approximated as
\[ N(0, \left( \frac{1}{n_1} + \frac{1}{n_2} \right) \times p(1 - p)), \]
where \( N \) designates the normal distribution. Under the condition that the null hypothesis \( H_0 \) is right, the common possibility \( p \) is estimated as
\[ \hat{p} = \frac{y_1 + y_2}{n_1 + n_2}, \]
In this case, \( \hat{p}_1, \hat{p}_2, \) and \( \hat{p} \) are calculated as
\[ \hat{p}_1 = \frac{8}{17} = 0.471, \]
\[ \hat{p}_2 = \frac{0}{5} = 0, \]
and
\[ \hat{p} = \frac{8}{22} = 0.364. \]

Then, a normalized statistics \( u \) is defined as
\[ u = \frac{\hat{p}_1 - \hat{p}_2}{\sqrt{\left( \frac{1}{n_1} + \frac{1}{n_2} \right) \times \hat{p}(1 - \hat{p})}}. \]

Under the null hypothesis \( H_0 \), the distribution of \( u \) can be approximated as the standard normal distribution \( N(0, 1) \). Here \( u \) is calculated as
\[ u = 1.92. \]  
\[ \Phi(u) = \int_{-\infty}^{u} \frac{1}{\sqrt{2\pi}} \exp \left[ -\frac{u'^2}{2} \right] du'. \]  
\[ \Phi(1.92) = 0.973. \]

In other words, under the condition that the null hypothesis \( H_0 \) is right, the possibility of \( u > 1.92 \) is only 2.7%. Therefore, the null hypothesis \( H_0 \) is rejected with a significant level of 97.3%. Therefore, it can be stated that "The weather does not tend to be fair on the day following a day of fair weather," when the summer atmosphere remains stable.

If all years from 1961 to 1996 are selected for the same analysis conditions, the total number of cases is 35. Figure 3 shows the results. In this case, the hypothesis is rejected with a significance level of 94.3%, which is smaller than that in the selected years.

### 3 Numerical experiments

#### 3.1 Basic equations

Numerical experiments were performed in order to examine the causes of the emergence of interdiurnal variations of cloud amount. Basic equations used in the present study is similar to those used by Yamasaki (1975). The main difference is that parameterizations of moist processes are simplified. Use is made of a two-dimensional model (yz-plane) which takes the Coriolis force into account. The effect of this force can be taken into account by introducing an x-directed wind perpendicular to the considered yz-plane. Non-acoustic conditions are assumed, but the hydrostatic balance is not assumed. The z-coordinate is taken as the altitude. Vorticity \( \xi \) can be used rather than the wind velocities \( v \) and \( w \) on the yz-plane, since non-acoustic conditions are assumed. The prognostic variables are the x-directed wind \( u \), the vorticity \( \xi \), the potential temperature \( \theta \), the water vapor mixing ratio \( q_v \), and the cloud water mixing ratio \( q_c \). The prognostic equations are written as

\[ \frac{D}{Dt} u = f v + \nu_H \frac{\partial^2 u}{\partial y^2} u + \frac{1}{\bar{\rho}} \frac{\partial}{\partial z} \left( \bar{\rho} \nu_v \frac{\partial}{\partial z} u \right), \]  
\[ \frac{D}{DT} \xi = f \frac{\partial}{\partial z} u + g \frac{\partial}{\partial y} \left( \frac{\theta}{\bar{\theta}} + 0.608 q_v - q_c \right) \]  
\[ + \nu_H \frac{\partial^2 \xi}{\partial y^2} + \frac{1}{\bar{\rho}} \frac{\partial}{\partial z} \left( \bar{\rho} \nu_v \frac{\partial}{\partial z} \xi \right), \]
\[ \frac{D}{DT} \theta = \frac{\bar{\theta}}{T} \frac{L}{C_p} C + \nu_H \frac{\partial^2}{\partial y^2} \theta + \frac{1}{\bar{\theta}} \left\{ \bar{\rho} \nu_V \frac{\partial}{\partial z} (\theta - \bar{\theta}) \right\} - \frac{1}{\tau_\theta} (\theta - \bar{\theta}), \]  

(21)

\[ \frac{D}{Dt} q_v = -C + \nu_H \frac{\partial^2}{\partial y^2} q_v + \frac{1}{\bar{\rho}} \left\{ \bar{\rho} \nu_v \frac{\partial}{\partial z} (q_v - \bar{q}_v) \right\}, \]  

(22)

and

\[ \frac{D}{Dt} q_c = C - R_d + \nu_H \frac{\partial^2}{\partial y^2} q_v + \frac{1}{\bar{\rho}} \frac{\partial}{\partial z} \left( \bar{\rho} \nu_v \frac{\partial}{\partial z} q_c \right), \]  

(23)

where

\[ \frac{D}{Dt} \equiv \frac{\partial}{\partial t} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z}. \]  

(24)

Here \( f \) represents the Coriolis parameter, \( g \) the acceleration of gravity, \( C_p \) the specific heat of air at constant pressure, \( L \) the heat of condensation of water per unit mass, \( \nu_H \) the coefficient of horizontal diffusion, \( \nu_V \) the coefficient of vertical diffusion, and \( \tau_\theta \) the time scale of Newtonian cooling. In the present study, these parameters are assigned to

\[ f = 7.27 \times 10^{-5} \ [\text{s}], \]  

(25)

\[ g = 9.81 \ [\text{m/s}^2], \]  

(26)

\[ C_p = 1003 \ [\text{J/kgK}], \]  

(27)

\[ L = 2.50 \times 10^6 \ [\text{J/kg}], \]  

(28)

\[ \nu_H = 1.0 \times 10^3 \ [\text{m}^2/\text{s}], \]  

(29)

\[ \nu_V = 5.0 \ [\text{m}^2/\text{s}], \]  

(30)

and

\[ \tau_\theta = 1.0 \times 10^6 \ [\text{s}]. \]  

(31)

Hereabove, \( \rho \) denotes the density, and \( T \) temperature of the air. Further, \( \bar{\theta}, \bar{q}_v, \bar{\rho}, \) and \( \bar{T} \) are the basic fields of \( \theta, q_v, \rho, \) and \( T \) depending only on \( z \), and are constant in time. Wind velocities in the \( yz \)-plane, \( v \) and \( w \), are calculated from the stream function \( \Psi \), which is in turn obtained from vorticity \( \xi \) by solving the differential equation. The equations for \( \Psi, v \), and \( w \) are
\[
\left( \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2} + \frac{2 d \rho}{\rho d z} \frac{\partial}{\partial z} + \frac{1 d^2 \rho}{\rho d z^2} \right) \Psi = \xi, \quad (32)
\]

\[
v = -\frac{1}{\rho} \frac{\partial}{\partial z} (\rho \Psi), \quad (33)
\]

and

\[
w = \frac{\partial}{\partial y} \Psi. \quad (34)
\]

The \( C \) in eqs. (21) - (23) represents the rate of condensation of water vapor into cloud water. It is assumed that any surplus water vapor condensates at a constant time scale \( \tau_c \) if the water vapor mixing ratio exceeds the saturated mixing ratio. On the other hand, cloud water evaporates at a constant time scale \( \tau_c \) if the atmosphere is not saturated. That is, \( C \) is defined as

\[
C = \begin{cases} 
\frac{1}{\tau_c} (q_v - q_v^*) & (q_v > q_v^*) \\
-\frac{1}{\tau_c} (q_v^* - q_v) & (q_v \leq q_v^*, q_c > 0) \\
0 & (q_v \leq q_v^*, q_c = 0)
\end{cases} \quad (35)
\]

Here, \( q_v^* \) designates saturated mixing ratio of water vapor, and is calculated as

\[
q_v^* = 0.622 \frac{e_s}{\rho - e_s}, \quad (36)
\]

where

\[
e_s = 611 \exp \left[ 17.27 \times \frac{T - 273.16}{T - 35.86} \right] \, [Pa] \quad (37)
\]

represents the saturated water vapor pressure and \( \rho \) the air pressure of the basic field. In the present study, \( \tau_c \) and \( \tau_c \) are assigned to

\[
\tau_c = 600 \, [s], \quad (38)
\]

and

\[
\tau_e = 600 \, [s]. \quad (39)
\]

The \( R_d \) in eq. (23) represents the conversion of cloud water into rain drops. It is assumed that surplus cloud water changes into rain drops at a constant time scale \( \tau_r \) if the cloud water mixing ratio exceeds a certain value \( q_{c_{max}} \), and that rain drops immediately precipitate. Therefore, \( R_d \) is defined as

\[
R_d = \begin{cases} 
\frac{1}{\tau_r} (q_c - q_{c_{max}}) & (q_c > q_{c_{max}}) \\
0 & (q_c \leq q_{c_{max}})
\end{cases} \quad (40)
\]
where
\[ q_{\text{e, max}} = 10^{-3} \text{ [s]}, \tag{41} \]
and
\[ \tau_r = 1200 \text{ [s]} . \tag{42} \]

### 3.2 Initialization of the basic field

The initial basic fields were set according to the climatological values for the Kanto plain in August. The surface pressure was set to the average of the climatological values at Tokyo, Kumagaya, and Maebashi (see Fig. 1) as
\[ \bar{p}(z = 0) = 1.0095 \times 10^5 \text{ [Pa]} . \tag{43} \]
Table 2 lists the height distributions of \( \bar{\theta} \) and \( \bar{q}_v \). The values of \( \bar{\theta} \) and \( \bar{q}_v \) at each height were calculated from the climatological values for the temperature and relative humidity at Tateno (see Fig. 1) except for \( \bar{\theta} \) and \( \bar{q}_v \) at the surface and \( \bar{q}_v \) at 500 hPa, 300 hPa, and 200 hPa. \( \bar{\theta} \) at the surface was calculated from the values of \( \bar{\theta} \) at 850 hPa and 700 hPa by linear extrapolation. \( \bar{q}_v \) at the surface was calculated from \( \bar{\theta} \) at the surface and the average of the climatological values of relative humidity at Tokyo, Kumagaya, and Maebashi. The values of \( \bar{q}_v \) above 500 hPa were set to 0 in order to avoid the formation of persisting stratiform clouds in the upper troposphere. The values of \( \bar{\theta} \) and \( \bar{q}_v \) at intermediate heights were obtained by linear interpolation. \( \bar{\rho} \) and \( \bar{\rho} \) at each height were obtained using the equation of state and the equation of hydrostatic balance, i.e.,
\[ \bar{\rho} = \bar{\rho} R \bar{\bar{\theta}}, \tag{44} \]
and
\[ \frac{d\bar{\rho}}{dz} = -\bar{\rho} g. \tag{45} \]
The temperature \( T \) was calculated from potential temperature \( \bar{\theta} \), using the following equation,
\[ \bar{T} = \bar{\theta} \times \left( \frac{\bar{p}}{p_0} \right)^{\frac{\gamma}{\kappa}} . \tag{46} \]
Here \( R \) is the gas constant and \( p_0 \) the standard pressure for the calculation of potential temperature. In the present study, \( R \) and \( p_0 \) are assigned to
\[ R = 287 \text{ [J/kgK]}, \tag{47} \]
and
\[ p_0 = 10^5 \text{ [Pa]} . \tag{48} \]
3.3 Boundary conditions

In order to solve eq. (32), boundary conditions $\Psi = 0$ are set at the top, bottom, and lateral boundaries, so that $w = 0$ at the top and bottom boundaries, and $v = 0$ at the lateral boundaries. Furthermore, at the top boundary, other boundary conditions are applied as

\[
\begin{align*}
  u &= 0, \quad (49) \\
  \xi &= 0, \quad (50) \\
  \theta &= \bar{\theta}, \quad (51) \\
  q_v &= \bar{q}_v, \quad (52) \\
  q_c &= 0, \quad (53)
\end{align*}
\]

and at the ground surface,

\[
\begin{align*}
  u &= 0, \quad (54) \\
  \xi &= 0, \quad (55)
\end{align*}
\]

and

\[
q_c = 0. \quad (56)
\]

The fluxes across the lateral boundaries of x-directed momentum, vorticity, sensible heat, water vapor, and cloud water are fixed to zero.

The surface boundary conditions for $\theta$ and $q_c$ are given as the sensible heat flux $F_{SH}$ and the latent heat flux $F_{LH}$ on the surface respectively, as

\[
F_{SH} = \alpha_{SH} \bar{C}_p \bar{\rho} \left[ \left\{ \theta_s - \bar{\theta} (z = 0) \right\} - \left\{ \theta (z = \frac{1}{2} \Delta z) - \bar{\theta} (z = \frac{1}{2} \Delta z) \right\} \right] \\
\times \frac{\bar{F}(z = 0)}{\bar{\theta}(z = 0)}. \quad (57)
\]

and

\[
F_{LH} = \alpha_{LH} L \bar{\rho} \left\{ q_v^* (z = 0) - q_v (z = \frac{1}{2} \Delta z) \right\}, \quad (58)
\]

where $\Delta z$ denotes the vertical grid interval. In the present study, $\alpha_{SH}$ and $\alpha_{LH}$ are assigned values for two different surfaces, i.e., the sea and land as follows,
\[ \alpha_{SH} = 0.025 \, [m/s] \quad \text{(sea and land)}, \] (59)

and

\[ \alpha_{LH} = \begin{cases} 0.025 \, [m/s] & \text{(sea)} \\ 0.0025 \, [m/s] & \text{(land)} \end{cases}. \] (60)

In order to avoid the accumulation of water vapor near the lateral boundaries where the motion of air is very small, \( F'_{LH} \) is used instead of \( F_{LH} \) within 20 km from the lateral boundaries. \( F'_{LH} \) is obtained as

\[ F'_{LH} = \begin{cases} 0 & (F_{LH} > 0) \\ F_{LH} & (F_{LH} \leq 0) \end{cases}. \] (61)

The value of \( \theta_s \) in eq. (57) is fixed to \( \bar{\theta}(z = 0) \) on the sea surface. On the other hand, \( \theta_s \) on the land surface is calculated as

\[ \theta_s = T_s \times \left( \frac{p_0}{\bar{p}(z = 0)} \right)^\frac{2}{7}, \] (62)

where \( T_s \) represents the surface temperature. The time derivation of \( T_s \) is obtained as

\[ C_{soil} \frac{\partial}{\partial t} T_s = F^\downarrow_S + F^\downarrow_L - F^\uparrow_L, \] (63)

where \( C_{soil}, F^\downarrow_S, F^\downarrow_L, \) and \( F^\uparrow_L \) represent the specific heat of the surface soil, insolation, downward long-wave radiation, and the upward long-wave radiation, respectively. In the present study, \( C_{soil} \) is

\[ C_{soil} = 8.0 \times 10^4 \, [J/Km^2]. \] (64)

\( F^\downarrow_S, F^\downarrow_L, \) and \( F^\uparrow_L \) and obtained as

\[ F^\downarrow_S = \begin{cases} 375 \sin \left[ \pi \left( \frac{t-t_1}{t_2-t_1} \right) \right] & [W/m^2] \quad (t_1 \leq t \leq t_2) \\ 0 & [W/m^2] \quad (t \leq t_1, t_2 \leq t) \end{cases}. \] (65)

\[ F^\downarrow_L = 330 \, [W/m^2], \] (66)

\[ F^\uparrow_L = \sigma T^4_s. \] (67)

where \( t \) and \( \sigma \) represent the local standard time and the Stefan-Boltzmann constant. \( t_1, t_2, \) and \( \sigma \) are

\[ t_1 = 0500 \text{LST}, \] (68)

\[ t_2 = 1900 \text{LST}, \] (69)

\[ \sigma = 5.67 \times 10^{-8} \, [W/m^2 K^4]. \] (70)
3.4 Standard experiment

3.4.1 Method

The size of the model region is 150 km horizontally and 10 km vertically. Sea surface is assigned to the region within 50 km from the left boundary \((y < 0)\) and land surface to the region within 100 km from the right boundary \((y > 0)\). The horizontal size is a little larger than the horizontal scale of the sea breeze over the Kanto plain. In order to avoid the occurrence of any convection which is not related to the sea breeze system, the size of the region is limited. The grid interval is \(\Delta y = 1000\) m horizontally and \(\Delta z = 200\) m vertically. The initial values equal those of the basic field. The model was integrated for 45 days, and the last ten days of data were analyzed. The time step was \(\Delta t = 36\) s. However, in order to save computation times, horizontal grid interval of \(\Delta y = 5000\) m and a time step of \(\Delta t = 120s\) were used for the first 30 days. In this coarse horizontal resolution case, the horizontal diffusion coefficient was set to

\[
\nu_H = 5.0 \times 10^3 \ [m^2/s],
\]

(71)

corresponding to the change in \(\Delta y\), so that the development of cumulus convection was similar to that in the case of \(\Delta y = 1000\) m. When the horizontal grid size was changed from \(\Delta y = 5000\) m to \(\Delta y = 1000\) m, the results of the model integration did not change to significantly.

3.4.2 Results

The sea breeze was simulated every day in the coastal region. The maximum wind speed was about 5 m/s, being consistent with observational results (Ogura, 1997). Figure 4 shows the diurnal variations of the sea surface temperature (solid line) and the land surface temperature averaged throughout the region of \(y (= 0 - 80 km)\) (dashed line). The variations were obtained by averaging the diurnal variations for the last 10 days of the integration. Note that the diurnal variations of surface temperature do not change day to day to any great extent. The maximum land surface temperature is about 1.8 °C higher than the average of the climatological values at the three cloud amount sites. On the other hand, the minimum land surface temperature is about 1.5 °C lower than the averaged climatological value. The diurnal variation of the land surface temperature shown in Fig. 4 can be thought to represent the actual diurnal variation on fine weather day. Figure 5 shows the time change of the values of the vertical column of cloud water in the region of \(y (= 0 - 80 km)\), after converting to units of precipitation. Day 0 in the figure denotes the 35th simulated day. Figure 6 shows the temporal distribution of precipitation per hour in the same region. Days with strong convection and those with weak convection appear quasi-periodically. Figure 7 shows the diurnal changes of the cloud water content shown in Fig. 5 on day 8 (solid line), and day 9 (dashed line). On both days, cumulus clouds
appear just after noon. But the strengths of convection are quite different day to day. Figures 8 and 9 show the wind and cloud distributions at 2000 LST on day 8 and day 9, respectively. On day 8, the growth of the cloud has already ended by 2000 LST, while a strong upward motion can be seen in the cloud which is still growing at 2000 LST on day 9. Figures 10 and 11 show the respective differences in potential temperature and water vapor mixing ratio at 0900 LST between day 8 and day 9. On day 9, potential temperatures are reduced around the altitude of 4 km, and the stability of the lower atmosphere is $0.10 - 0.15$ K/km less than that on day 8. Furthermore, the water vapor mixing ratio on day 9 is $0 - 0.35$ g/kg greater than that on day 8 in the lower and middle atmosphere. Figure 12 shows the surface heat fluxes in the region of $y (= 0 - 80 km)$. Both the sensible (solid line) and latent (dashed line) heat fluxes change diurnally. Note here that interdiurnal variations cannot be detected. The strength of convection changes day to day, although the surface heat fluxes indicate almost the same diurnal changes everyday. This means that interdiurnal variations of cumulus convection appear as an internal mode of the atmosphere. Interdiurnal variations cannot be seen in the experiment without any moist processes. Therefore, it can be said that interdiurnal variations are an internal mode caused by moist convection.

### 3.5 Comparative experiments

To analyze how the differences in potential temperature and water vapor mixing ratio influence cumulus convection, several comparative experiments were conducted. In experiment 1, the model was integrated for 24 hours using the initial conditions at 0900 LST of day 8. The experiment generated the same results as the original experiment on day 8. In experiment 2, only the values of water vapor mixing ratio on day 8 were replaced with those at 0900 LST on day 9 of the original experiment. In a similar manner in experiment 3, only the values of potential temperature at 0900 LST on day 8 were changed to those on day 9. In experiment 4, both the values of the mixing ratio and potential temperature on day 8 were replaced with those on day 9.

The result is shown in Figure 13. When either the values of mixing ratio (dashed line) or potential temperature (dotted line) are changed, cumulus convection becomes stronger. When both of the values of mixing ratio and potential temperature are replaced, the convection becomes stronger still. Therefore it can be concluded that interdiurnal variations of cumulus convection are caused by the changes in both the mixing ratio and the potential temperature. Note here that cumulus convection is stronger when the values of potential temperature are modified than when the values of mixing ratio are. This does not mean that water vapor does not influence interdiurnal variations compared to sensible heat, because interdiurnal variations can not be seen in dry experiment. Actually, it can be stated that the change in the potential temperature is caused by the moist processes.
3.6 Sensitivity experiments

Several sensitivity experiments were performed to examine the characteristics of interdiurnal variations found in the standard experiment. In experiment A and B, only the basic field of potential temperature is changed. In experiment A, the vertical gradient of potential temperature was reduced by 0.5 K/km than that of the standard experiment, without changing the value at \( z = 0 \). In experiment B, the vertical gradient was increased by 0.5 K/km. Figures 14 and 15 show the time variations of cloud water content of experiments A and B, respectively. In experiment A, the cumulus convection is stronger and interdiurnal variations can be easily detected to some extent. In experiment B, on the other hand, the convection is weaker and interdiurnal variations are not as pronounced as in experiment A.

Other experiments were run in which the surface conditions of the latent heat flux have been altered. In experiment C, the values of \( \alpha \) in eq. (60) were changed as

\[
\alpha_{LH} = \begin{cases} 
0.025 & \text{[m/s]} \quad \text{(sea)} \\
0 & \text{[m/s]} \quad \text{(land),}
\end{cases}
\]  

(72)

while in experiment D,

\[
\alpha_{LH} = \begin{cases} 
0.025 & \text{[m/s]} \quad \text{(sea)} \\
0.005 & \text{[m/s]} \quad \text{(land).}
\end{cases}
\]  

(73)

Experiment C corresponds to a weaker latent heat flux than that of the standard experiment. Experiment D, on the other hand, corresponds to a stronger latent heat flux. Figures 16 and 17 show the time change of cloud water content in experiments C and D, respectively. In experiment C, as expected, the strength of cumulus convection does not change appreciably and interdiurnal variations can be easily detected. It means that water vapor is supplied mainly from the sea surface in the standard experiment. In experiment D, however, the convection is stronger and interdiurnal variations cannot be seen clearly. From the results of the sensitivity experiments, it can be stated that interdiurnal variations of cumulus convection readily appear when the averaged strength of cumulus convection has a medium value. In experiment D, it seems that the strength of convection is saturated because the growth of convection is limited even if more water vapor is supplied. However, such a limitation may be unrealistic because the interdiurnal variation can be seen in a much wetter condition such as those over the tropical ocean (Chen and Houze, 1997). Another model which can simulate large convective systems is needed to examine the interdiurnal variation of such large convection which can be seen over the tropical ocean.

Further experiments were performed with less or no diurnal variation of surface temperature corresponding to oceanic conditions (cf. Chen and Houze, 1997). In experiment E, specific heat of the surface soil \( C_{soil} \) was changed as
\[ C_{\text{soil}} = 8.0 \times 10^5 \text{ [J/K-m}^2\] \] (74)

Figure 18 shows the averaged diurnal variation of the surface temperature over days 1 – 10 throughout the region of \( y (= 0 \text{ – 80km}) \). The diurnal range of the surface temperature is 5.8 °C, while that in the standard experiment is 11.4 °C. Figure 19 shows the time change of cloud water content in experiment E. Interdiurnal variations can be detected as in the standard experiment. At last, experiment F were performed in which the sea surface condition was set for the whole region except for the first 30 days. For the first 30 days, the model was integrated in exactly the same way as in the standard experiment in order to make the horizontal uniformity arise. Figure 20 shows the result. Convective events with the same strength arise at a constant interval. Interdiurnal variations are not apparent.

It can be stated that interdiurnal variations can appear under the condition of narrower diurnal range of surface temperature, but does not appear in the absence of diurnal range. It seems that a periodic forcing is important to interdiurnal variations. Note that the interval of cumulus convection is about 18 hours. This is consistent with the fact that a time of about 18 hours is required for the mixed layer to recover to previous conditions over the tropical ocean (Chen and Houze, 1997). However, under a narrower moist condition such as over the land surface, the recovery time of cumulus convection is somewhat longer than over the tropical ocean.

Furthermore, another experiment was performed by setting the Coriolis parameter \( f \) to zero. The results, however, exhibited only slight changes. It has been demonstrated that the Coriolis force does not influence interdiurnal variations.

### 3.7 Step experiments

In order to examine the characteristics of the interdiurnal variations, several 1-day integrations with initial conditions at 0900 LST were performed. Here, not only conditions at 0900 LST for days with strong and weak convection but also some intermediate conditions between them were used as initial conditions. First, 1-day integrations were operated for the result of experiment E in the previous section. The x-directed wind \( u \) was set to

\[
    u_0(y, z) = \frac{1}{2} \left\{ u^+(y, z) + u^-(y, z) \right\} + \frac{1}{2} \beta_0 \left\{ u^+(y, z) - u^-(y, z) \right\},
\]  

(75)

where \( u^+ \) and \( u^- \) represent the values of \( u \) at 0900 LST on day 6 when convection is weak and day 7 when convection is strong, respectively. \( \beta_0 \) is constant. The initial conditions of \( \xi \), \( \theta \), \( q_e \), \( q_v \), and \( T_s \) are set in the same manner as \( u_0(y, z) \). The values of \( \beta_0 \) are given for respective integrations as

\[
    \beta_0 = -2.50 + 0.25(n - 1)(n = 1, 2, ..., 21).
\]  

(76)
After 1-day integration, the value of \( \beta_1 \) is determined so that the following integration over all the model has a minimum value.

\[
I = \int \int [\theta_1(y, z) - \frac{1}{2} \left\{ \theta^+(y, z) + \theta^-(y, z) \right\} + \frac{1}{2} \beta_1 \left\{ \theta^+(y, z) - \theta^-(y, z) \right\}]^2 \, dy \, dz, \quad (77)
\]

where \( \theta_1 \) represents the value of \( \theta \) after 1-day integration. Note that only the values of \( \theta_1 \) are used to obtain \( \beta_1 \). This is justified by the fact that the variance of \( \theta \) affects more largely on the diurnal variations than that of \( q_v \) in the comparative experiments presented in subsection 3.5. Even if the values of \( q_v \) are used instead of those of \( \theta \), the estimation of \( \beta_1 \) does not change significantly. Figure 21 shows \( \beta_1 \) in function of \( \beta_0 \). As a matter of course, \( \beta_1 = 1 \) when \( \beta_0 = -1 \). If a complete 2-day periodicity could be assumed, then \( \beta_1 = -1 \) when \( \beta_0 = 1 \). In the range of \(-1 \leq \beta_0 \leq 1.25 \), \( \beta_1 \) decreases as \( \beta_0 \) increases. It means that the more favorable the condition is for cumulus convection on a given day, the less favorable it will become for another cumulus convection on the following day as a result of greater convection on the first day. The line \( \beta_0 = f(\beta_1) \) is also drawn for \( \beta_1 = f(\beta_0) \). The intersections of the two lines are indicated as \( A(\beta_A, \beta_B) \), \( B(\beta_B, \beta_A) \), and \( C(\beta_C, \beta_C) \). \( \beta = \beta_C \) may be a static solution, because the value of \( \beta \) does not change after 1-day integration. However, the value of the derivative function \( f'(\beta) \) at \( \beta = \beta_C \) does not satisfy the following condition:

\[
|f'(\beta)| < 1. \quad (78)
\]

Therefore, the solution \( \beta = \beta_C \) is unstable. On the other hand, \( f'(\beta) \) satisfies the following condition concerning \( \beta_A \) and \( \beta_B \).

\[
|f'(\beta_A) \times f'(\beta_B)| < 1. \quad (79)
\]

The repetition of \( \beta = \beta_A \) and \( \beta = \beta_B \) can arise rather than the stable solution \( \beta = \beta_C \). In the above, it is assumed that the situation of the fields of all prognostic variables can be represented by one parameter \( \beta \) for simplicity. The degree of freedom in the studied case, is lower than that in the real atmosphere or in a numerical experiment. The degree of freedom being indeed high makes the behavior of interdiurnal variation in a numerical experiment quite complex.

Similar 1-day integrations were performed for the result of the standard experiment with initial conditions composed from the values of the prognostic variables at 0900 on day 8 and day 9. In Figure 22, the interdiurnal variations are a little difficult to explain by taking into account only the relation linking between \( \beta_0 \) and \( \beta_1 \), although they have a similar behavior as in Fig. 21. A remark can be made that the variation which is not represented by the variable \( \beta \) also plays an important role in the interdiurnal variation.
4 Analysis of the aerological data

The aerological data of temperature and humidity at Tateno (see Fig. 1) were examined in comparison with the numerical results shown in Figs. 10 and 11. Data from this site were selected as representative of the three cloud amount sites. The same years and same period of the year were chosen as defined in section 2. First, the mean values of cloud amount at the three sites for every morning and afternoon were obtained in the same manner as in section 2. If the mean cloud amount values were less than 4 during the mornings of three consecutive days, an examination was made of the mean afternoon cloud amount, and the vertical profiles of potential temperature and water vapor mixing ratio at 0900 LST both for the first day and the second day. The values of potential temperature and mixing ratio at each level were averaged for cases when the mean cloud amount in the afternoon was greater than 5 and less than 5.

Table 3 lists the results for potential temperature (the upper line in each column) and water vapor mixing ratio (the lower line in each column) at each level. In addition, the vertical gradients of potential temperature between the surface and 500 hPa were calculated (the right-hand column). The unbiased standard deviations of the potential temperature gradient and the mixing ratio are also listed. For days with large cloud amount in the afternoon, the lower atmosphere is about 0.17 K / km less stable than that on less cloudy days, and the mixing ratios in the lower and middle atmosphere are from 0.2 to 2.0 g/kg greater in the mornings.

Here, t-testings were performed in order to examine the significance of the tendency mentioned above. First, the significance of the difference of the vertical gradient of potential temperature is tested. The null hypothesis $H_0$ and the alternative hypothesis $H_1$ can be set as follows.

$$H_0 : \mu_1 = \mu_2,$$

$$H_1 : \mu_1 < \mu_2,$$

where $\mu_1$ and $\mu_2$ represent the mean values of the vertical gradient of potential temperature in the cases in which the cloud amount is greater than or equal to 5 and less than 5, respectively. In this case, $\mu_1$ and $\mu_2$ are estimated as

$$\tilde{\mu}_1 = 4.44,$$

$$\tilde{\mu}_2 = 4.61.$$

Then, the difference of the estimated values is calculated as

$$\tilde{\mu}_1 - \tilde{\mu}_2 = -0.17.$$
On the other hand, under the null hypothesis $H_0$, the expected value and the deviation of $\bar{\mu}_1 - \bar{\mu}_2$ are

$$E(\bar{\mu}_1 - \bar{\mu}_2) = 0,$$

(85)

and

$$V(\bar{\mu}_1 - \bar{\mu}_2) = V(\bar{\mu}_1) + V(\bar{\mu}_2) = \frac{s_1^2}{n_1} + \frac{s_2^2}{n_2},$$

(86)

where $n_1$ and $n_2$ are the numbers of cases in which the cloud amount is greater than or equal to 5 and less than 5, while $s_1$ and $s_2$ are the population standard deviations in the respective cases. In this analysis,

$$n_1 = 11,$$

(87)

$$n_2 = 27,$$

(88)

$$s_1 = 0.41,$$

(89)

$$s_2 = 0.44.$$

(90)

Here, it is assumed that $s_1$ and $s_2$ are equal to the pooled standard deviation $s$ obtained from $s_1$ and $s_2$. Then, $s$ can be calculated as

$$s = \sqrt{\frac{(n_1 - 1)s_1^2 + (n_2 - 1)s_2^2}{n_1 + n_2 - 2}} = 0.43.$$

(91)

The deviation of $\bar{\mu}_1 - \bar{\mu}_2$ can be calculated as

$$V(\bar{\mu}_1 - \bar{\mu}_2) = s^2 \left( \frac{1}{11} + \frac{1}{27} \right) = 0.02366.$$

(92)

Then,

$$\sqrt{V(\bar{\mu}_1 - \bar{\mu}_2)} = 0.154.$$

(93)

Here, the t-statistics is introduced as follows.

$$t = \frac{\bar{\mu}_1 - \bar{\mu}_2}{\sqrt{V(\bar{\mu}_1 - \bar{\mu}_2)}}$$

(94)

In this case, $t$ is calculated as

$$t = \frac{-0.17}{0.154} = -1.10.$$  

(95)
Under the null hypothesis $H_0$, the distribution of the t-statistics is expected to be the t-distribution with degree of freedom $n_1 + n_2 - 2$. Here,

$$t_{0.05}(n_1 + n_2 - 2) = t_{0.05}(36) = 1.69,$$

and then,

$$|t| < t_{0.05}(36).$$

Therefore, the null hypothesis $H_0$ can not be rejected with a significance level of 95%. It can be concluded that the difference of the vertical gradient of the potential temperature is not significant. Then, the significance of the difference of the mixing ratio at 700 hPa is tested in the same manner. Here, the null hypothesis $H'_0$ and the alternative hypothesis $H'_1$ can be set as

$$H'_0 : \mu'_1 = \mu'_2,$$

and

$$H'_1 : \mu'_1 > \mu'_2,$$

where $\mu'_1$ and $\mu'_2$ represent the expected values of the mixing ratio at 700 hPa in the respective cases. The t-statistics is calculated as

$$t' = \frac{2.0}{0.773} = 2.59,$$

and then,

$$|t'| > t_{0.05}(36).$$

Therefore, the null hypothesis $H'_0$ can be rejected with a significance level of 95%. It is shown that the difference of the mixing ratio is significant. It can be concluded that only the difference of the mixing ratio is significant, although those of both the potential temperature gradient and the mixing ratio are qualitatively consistent with the results of the numerical experiments presented in the previous section.

Note here that the differences are less at the surface compared to those in the free atmosphere. It means that the changes in the surface conditions such as temperature and wetness do not play any important role. Rather, the changes in the free atmosphere such as those in cumulus convection, are important. It is not clear, however, that all of the differences in the potential temperature and the mixing ratio obtained from the aerological data, resulted from cumulus convection which had occurred until the previous day. In fact, the differences taken from the aerological data are larger than those found in the results of the numerical experiments. The conditions of the lower atmosphere may be influenced by changes in the large-scale field, which are not accounted for in the numerical experiments, but may be present in the observed data.
5 Conclusions

By analyzing cloud amount data over the Kanto plain in Japan, a negative correlation has been found between the cloud amount in the afternoon of a given day and that of the following, when the summer atmosphere is stable. This correlation was found to be statistically significant.

According to the results of numerical experiments, the mechanism of interdiurnal variations of cumulus convection over the Kanto plain in Japan can be understood in the following way. It is assumed that the lower atmosphere is initially stable and the water vapor mixing ratio is small during the morning of a given day. In this case, cumulus convection does not develop significantly during the afternoon. Latent heat is not transported upward to any extent from the lower layer, and the amount of water vapor is not exhausted very much. The water vapor and sensible heat supplied from the surface are retained in the lower layer. By the next morning, the atmosphere has become more unstable and the mixing ratio has increased. Cumulus convection can now develop to a greater extent in the afternoon. Much of the latent heat is transported upward, and water vapor is largely exhausted, even when the sensible and latent heat fluxes at the surface are relatively small. As a result, conditions unfavorable for cumulus convection have been set up again for the next morning. This on and off development of conditions for convection results in the two-day cumulus convective oscillation. The effect of the Coriolis force was found to be very small on the interdiurnal variations.

In the numerical experiments, changes in insolation at the surface and in the radiation in the atmosphere caused by cumulus cloud were not considered. According to Chen and Houze (1997), these radiative effects are important to the large convective system with the cloud deck lasting even until the day following a convective event. However, in the present study, cumulus convection which is not so large and does not last until the next morning was considered. Furthermore, the analysis of aerological data presented in the previous section shows that the changes in potential temperature and mixing ratio are greatest in the free atmosphere. Therefore, it can be stated that radiative effects do not affect the interdiurnal variations to any great extent.

According to the results of sensitivity experiments (Figs. 14 - 17, and 19), interdiurnal variations generally appear to a certain degree. However, in the analysis of cloud amount data in section 2, the total number of cases is rather small. It can be said that the interdiurnal variation, which is clearly found in the numerical experiments, is not observed as often in the real atmosphere. This is thought to be due to the influence of large changes in the large-scale fields. In the aerological data (Table 3), those large-scale changes are greater than those seen in the numerical experiments. These data are almost certainly influenced by the large-scale fields, resulting in a less obvious interdiurnal oscillation. Furthermore, cumulus convection over the Kanto plain is affected by the topography such as mountains located at the north of the plain. Three-dimensional experiments with the
actual topography are required to estimate interdiurnal variations qualitatively.

The interdiurnal variation of cumulus convection examined in the present study is similar to the quasi two-day period oscillations of cumulus convection found over the tropical oceans, as mentioned in section 1. A long time (~18 hours) is required for the mixed layer to return to previous conditions (Chen and Houze, 1997). However, the geographical constraint over the tropical ocean is small, and diurnal variations of SST are small, while the influence of local circulations, such as sea and land breezes are large, and the surface temperature undergoes large changes in the coastal area. At this point, both oscillations are quite different from each other. In addition, the surface latent heat flux is much larger than the surface sensible heat flux over the tropical ocean (Lin and Johnson, 1996), while both fluxes have nearly the same magnitude over land surfaces. The cumulus convection over the coastal area is geographically strongly confined by sea and land breezes, so that the phase lock to westward-propagating gravity waves found in cumulus convection over the tropical ocean, can not be detected.

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The GFD-DENNOU Library was utilized for the drawing of the figures.

References


Figure 1: Locations of the observational sites in the Kanto plain. Cloud amount data from Tokyo, Kumagaya, and Maebashi were examined, while Tateno supplied aerological data.
Figure 2: Relationships between the mean cloud amount during the afternoon of a given day and that on the following day for selected years. The cloud amount values on the first day are plotted along the abscissa versus those on the second day on the ordinate. See the text for criterion used in selecting the data analyzed. Data analyzed consist of nine summers between 1961 and 1996.
Figure 3: The same as in Fig. 2, expect when selecting the summers of all of the years between 1961 and 1996.
Figure 4: Diurnal variation of the surface temperature $T_s$ [°C]. The solid line represents the sea surface temperature and the dashed line the land surface temperature averaged in the region of $y$ (= 0 - 80 km).
Figure 5: Temporal change of the value of the vertical column of cloud water in the region of $y (= 0 - 80 \text{km})$ over the last 10 days of model integration in the standard experiment. The values shown have been converted into precipitation units [mm].
Figure 6: The same as Fig. 5, except for the rate of precipitation [mm/hr].
Figure 7: Diurnal variation from 0900 LST of the day to 0900 LST of the following day of the cloud water content [mm] shown in Fig. 5 for day 8 (solid line) and day 9 (dashed line).
Figure 8: Time distributions of cloud and the wind in the yz-plane of the model at 2000 LST on day 8. The surface in the region of $y < 0$ is sea and $y > 0$ land. The hatched area represents the cloud, and the thick-hatched area the region where cloud water content is greater than 0.5 g/kg. The arrows indicate the wind velocity field. The wind scale is shown to the right of the figure.
Figure 9: The same as in Fig. 8, expect on day 9.
Figure 10: The difference in the values of potential temperature at 0900 LST between day 8 and day 9. The contour interval is 0.05 K. The region where potential temperature is less on day 9 is hatched.
Figure 11: The difference of water vapor mixing ratio at 0900 LST between day 8 and day 9. The contour interval is 0.05 g/kg. The region where water vapor mixing ratio is less on day 9 is hatched.
Figure 12: Temporal change of the surface heat fluxes [W/m²] averaged along the y-axis (0–80 km, land surface) over the last 10 days of the standard experiment. The solid indicates the sensible heat flux, and the dashed line the latent heat flux.
Figure 13: Diurnal variations of the cloud water content [mm] in the region of $y (= 0–80\text{km})$ over the whole period of model integration in the comparative experiments. The solid line indicates the results of experiment 1, the dashed line experiment 2, the dotted line experiment 3, and the dashed-and-dotted line experiment 4.
Figure 14: The same as in Fig. 5, except for experiment A.
Figure 15: The same as in Fig. 5, except for experiment B.
Figure 16: The same as in Fig. 5, except for experiment C.
Figure 17: The same as in Fig. 5, except for experiment D.
Figure 18: The same as in Fig. 4, except for experiment E.
Figure 19: The same as in Fig. 5, except for experiment E.
Figure 20: The same as in Fig. 5, except for experiment F.
Figure 21: Relationships between $\beta_0$ and $\beta_1$ for experiment E. The values of $\beta_0$ are plotted along the abscissa versus those of $\beta_1$ on the ordinate. See the text for the definitions of $\beta_0$ and $\beta_1$. The lines of $\beta_1 = f(\beta_0)$ and $\beta_0 = f(\beta_1)$ are also drawn. The marks of A, B, and C indicate the intersections of $\beta_1 = f(\beta_0)$ and $\beta_0 = f(\beta_1)$. 
Figure 22: The same as in Fig. 21, except for the standard experiment. Note that the definitions of $\beta_0$ and $\beta_1$ are slightly different from those in Fig. 21 concerning on the dates when the initial conditions are taken from.
<table>
<thead>
<tr>
<th></th>
<th>$C_2 &lt; 5$</th>
<th>$C_2 \geq 5$</th>
<th>total</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C_1 &lt; 5$</td>
<td>9</td>
<td>8</td>
<td>17</td>
</tr>
<tr>
<td>$C_1 \geq 5$</td>
<td>5</td>
<td>0</td>
<td>5</td>
</tr>
<tr>
<td>total</td>
<td>14</td>
<td>8</td>
<td>22</td>
</tr>
</tbody>
</table>

Table 1: Contingency table for the result shown in Fig. 2. $C_1$ and $C_2$ represent the afternoon cloud amounts on the first day and the second day, respectively.

<table>
<thead>
<tr>
<th>Pressure [hPa]</th>
<th>Altitude [m]</th>
<th>Potential temp. [K]</th>
<th>Mixing ratio [g/kg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>1009.5</td>
<td>0</td>
<td>298.9</td>
<td>15.7</td>
</tr>
<tr>
<td>850</td>
<td>1503</td>
<td>305.3</td>
<td>12.2</td>
</tr>
<tr>
<td>700</td>
<td>3145</td>
<td>313.3</td>
<td>7.1</td>
</tr>
<tr>
<td>500</td>
<td>5868</td>
<td>327.3</td>
<td>0.0</td>
</tr>
<tr>
<td>300</td>
<td>9702</td>
<td>343.1</td>
<td>0.0</td>
</tr>
<tr>
<td>200</td>
<td>12463</td>
<td>351.7</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Table 2: Height profiles of the basic fields of potential temperature [K] and water vapor mixing ratio [g/kg]. The values at intermediate heights are obtained by linear interpolation.

<table>
<thead>
<tr>
<th>Mean Altitude [m]</th>
<th>Number of Samples</th>
<th>Surface</th>
<th>850 hPa</th>
<th>700 hPa</th>
<th>500 hPa</th>
<th>$d\theta/dz$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>0</td>
<td>1503</td>
<td>3145</td>
<td>5868</td>
<td>4.56±0.43</td>
</tr>
<tr>
<td>All the Cases</td>
<td>38</td>
<td>299.8</td>
<td>306.2</td>
<td>314.3</td>
<td>326.6</td>
<td>4.44±0.41</td>
</tr>
<tr>
<td></td>
<td></td>
<td>16.8±2.4</td>
<td>9.1±2.9</td>
<td>4.9±2.3</td>
<td>1.5±0.8</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>17.1±2.5</td>
<td>9.7±3.4</td>
<td>6.3±2.3</td>
<td>1.6±0.7</td>
<td></td>
</tr>
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Table 3: Relation between the mean cloud amount in the afternoon and the aerological data at 0900 LST at the surface, 850 hPa, 700 hPa, and 500 hPa. Potential temperature [K] is shown above and water vapor mixing ratio [g/kg] below. The standard deviations are shown after ±. Unit of the vertical gradient of potential temperature is K/km.