Fine Structure of Vertical Motion in the Stratiform Precipitation Region Observed by a VHF Doppler Radar Installed in Sumatra, Indonesia

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ABSTRACT
Vertical motion $W$ profiles in the stratiform precipitation region of mesoscale cloud clusters were investigated using wind data observed by VHF Doppler radar installed in western Sumatra Island (0.2°S, 100.32°E). A special mode for $W$ observations was introduced in November 2003, and $W$ data with high accuracy were obtained during most of the period, with fine resolutions of 3 min in time and 150 m in vertical direction. The typical fine structure of $W$ within the nimbostratus in the stratiform precipitation region was investigated by the case study of 6, 8, and 20 November 2003. In the later 2 or 3 h of the stratiform precipitation period, gentle upward motions with small time and height fluctuations were observed over a several-kilometer height range from the middle to upper troposphere. Values of $W$ were weakly positive (0–40 cm s$^{-1}$) continuously, with little strong upward motion greater than 40 cm s$^{-1}$ and downward motion.

1. Introduction
The VHF Doppler radar installed in western Sumatra, Indonesia (0.2°S, 100.32°E) has enabled the observation of vertical motion $W$ with fine time (3 min) and vertical (150 m) resolutions. This paper describes the fine structure of $W$ within the stratiform precipitation region of tropical mesoscale cloud clusters.

In a mesoscale cloud cluster, the stratiform precipitation region, mainly consisting of nimbostratus, occupies most of the area. Vertical motions are related closely to dynamics and physics within nimbostratus. A comprehensive discussion of the dynamics and physics is found in chapter 6 of Houze (1993). Upward motion in nimbostratus can be induced by the following mechanisms: 1) old convective cells that still maintain buoyancy, even in the stratiform precipitation region; 2) diabatic heating through the phase change of water substances (e.g., Houze 1989; Hobbs et al. 1980); 3) gravity wave excitation in the convective region and propagation into the stratiform region (e.g., Pandya and Durran 1996); and 4) radiative effect (e.g., Churchill and Houze 1989).

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1991). On the other hand, the magnitude of upward motion affects the size and lifetime of nimbostratus, because it sustains precipitation particles within the clouds.

Many previous papers studied the representative values of $W$ within nimbostratus. Houze (1989) collected the results of past observational studies and demonstrated that upward motion in the stratiform precipitation region of tropical mesoscale cloud clusters was about $10-20 \text{ cm s}^{-1}$. However, the vertical profile of $W$ shown in most of the previous studies was computed only by indirect methods such as the vertical integration of horizontal divergence observed by radiosonde or weather radar.

Although indirect methods for the computation of $W$ can obtain representative values in the stratiform precipitation region, these methods have been shown to have difficulties in describing the fine structure of $W$ in tropical nimbostratus, as mentioned below. Estimation of $W$ from radiosonde data is limited to a large-area average and cannot detail the structure within the clouds. Weather radar cannot estimate $W$ without precipitation particles, even inside clouds. Besides, the method needs to hypothesize the boundary condition when integrating horizontal divergence vertically.

On the other hand, VHF Doppler radar can directly observe $W$ above the radar. VHF Doppler radar receives an echo from atmospheric turbulence with a scale of several meters (one-half of its wavelength), in both clear and cloudy regions (Fukao et al. 1985). There have been several studies on the distribution of $W$ in the stratiform precipitation region using VHF Doppler radars installed in the Pacific Ocean region (Gage et al. 1991). Balsley et al. (1988) presented statistical vertical profiles of $W$ for clear, stratiform rainfall and convective rainfall conditions at Ponape Island ($7^\circN, 157^\circE$) in the western Pacific. Cifelli and Rutledge (1994, 1998) and Cifelli et al. (1996) reported on the observational results of $W$ in northern Australia ($12.5^\circS, 131^\circE$) using VHF Doppler radar along with weather radar. The VHF Doppler radar located in Gadanki, India, ($13.5^\circN, 79.2^\circE$) was also used to observe $W$ in mesoscale convective systems (e.g., Kumar et al. 2001). They showed statistics of the vertical distribution of $W$ and several examples of the distribution of $W$ in regions such as the western and central Pacific, northern Australia, and India.

These results act as an inspiration for the current study to obtain further systematic descriptions of the fine structure of $W$ in nimbostratus. It is of interest to detail the time–height structure inside the clouds within the stratiform precipitation region with finer vertical and time resolutions than those of previous studies. Though regional differences in the basic state of the atmosphere such as wind shear, surface conditions, and cloud/precipitation particle size distribution may possibly cause differences in $W$ profiles within stratiform clouds, the number of the cases already reported is only a few and is not sufficient for a discussion on the effect of various basic states. It is worthy to conduct new $W$ observations in the Indonesian “Maritime Continent,” because it is located far from the locations of previous studies and has unique features in the cloud systems, such as diurnal variation (e.g., Mori et al. 2004) and a hierarchy of the clusters (e.g., Shibagaki et al. 2006).

Although VHF Doppler radar has the advantage of observing $W$ directly, the observation of $W$ in the upper troposphere at times proves difficult. The reflectivity from the upper-tropospheric region is generally weak, because of the weaker Bragg scattering echo than that in the lower and middle troposphere and the weak partial reflection that is dominant in the lower stratosphere. To overcome this problem, an operational mode of the Equatorial Atmospheric Radar (EAR) was arranged to improve the signal-to-noise ratio (SNR) of the Doppler spectra through use of a vertically pointing radar beam, as shown in the next section. As a result, the fine distributions of $W$ were observed over most of the entire height region in nimbostratus, with high time and vertical resolutions.

The data utilized are described—in particular, the handling technique of the EAR data—in section 2. The results are documented in section 3. A discussion is presented in section 4. Conclusions are given in section 5.

2. Data description

As was previously mentioned, VHF Doppler radar can directly observe $W$ and horizontal motions in the troposphere and lower stratosphere (e.g., Röttger 1980; Gage 1990). The EAR is a VHF Doppler radar operated at a frequency of 47 MHz and a peak output power of 100 kW. It was installed at the Equatorial Atmosphere Observatory at Kototabang ($0.2^\circS, 100.32^\circE$, 865 m MSL), which is located in the mountainous region of western Sumatra. For details of the EAR system, see Fukao et al. (2003).

During November of 2003, the EAR was operated with an additional vertical observation mode (hereinafter referred to as the vertical wind mode) to monitor $W$ intensively. Table 1 lists the observational parameters of the EAR. This mode and the normal observational mode were alternately carried out. Because the half-power full width of the EAR beam is 2.4°, the size of the sample volume becomes about 300 m at 8-km altitude. Radar beams were steered on a pulse-to-pulse
basis in both observation modes. To improve the observation range and accuracy of $W$, the SNR of Doppler spectra was improved. In the standard observation mode, the time for scanning in the vertical direction is about 16.4 s (=81.92/5 s). On the other hand, in the vertical wind mode all of the radar beams are pointed vertically in the observation time of 78.6 s to integrate more pulses than the standard observation mode; it contributes an SNR improvement of 7 dB for Doppler spectra obtained with the vertically pointed radar beam.

Values of $W$ are computed using the vertical wind mode. The following offline signal processing was carried out to derive $W$. First, three Doppler spectra, obtained simultaneously by three vertically pointing beams, which are switched every transmission, were averaged. Second, offline incoherent integrations were applied using four successive spectra in the time domain to reduce fluctuations in the Doppler spectra for deriving 12-min-resolution data. Though $W$ computation using four successive spectra has a time resolution of about 12 min, start times for $W$ computation are selected every 6 min to smooth $W$ profiles. No incoherent integrations were applied to 3-min resolution data. Third, the values of $W$ were estimated by assuming a Gaussian distribution of the atmospheric echo and applying a least squares fitting method to the Doppler spectra. For details of the application of the fitting method to the Doppler spectra, see Yamamoto et al. (1988). Echoes from refractivity fluctuations by Bragg scattering and those from precipitation particles by Rayleigh scattering are separated when a least square fitting method is applied. Last, the quality of $W$ was checked both automatically and manually to remove erroneous data resulting from the interference of signals and unreliable data in low signal-to-noise levels.

If Eq. (13) of Yamamoto et al. (1988) is applied, the estimated error in $W$ is 2.6 cm s$^{-1}$, even in considerably turbulent conditions for a Doppler width of 0.5 m s$^{-1}$ in the stratiform precipitation region. This accuracy is sufficient when observing the expected value of $W$ in nimbostratus (10–20 cm s$^{-1}$).

**TABLE 1. Parameters of EAR observation in standard routine observations and a special observation in November 2003.**

<table>
<thead>
<tr>
<th>Item</th>
<th>Standard mode</th>
<th>Vertical wind mode</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vertical resolution</td>
<td>150 m</td>
<td>150 m</td>
</tr>
<tr>
<td>Pulse repetition frequency</td>
<td>2500 Hz</td>
<td>2500 Hz</td>
</tr>
<tr>
<td>Beam direction (azimuth, zenith)</td>
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<td>(0°, 0°)</td>
</tr>
<tr>
<td></td>
<td>(0°, 10°)</td>
<td>(0°, 0°)</td>
</tr>
<tr>
<td></td>
<td>(90°, 10°)</td>
<td>(0°, 0°)</td>
</tr>
<tr>
<td></td>
<td>(180°, 10°)</td>
<td>(270°, 10°)</td>
</tr>
<tr>
<td>Total No. of coherent integrations</td>
<td>32</td>
<td>128</td>
</tr>
<tr>
<td>Total No. of FFT points</td>
<td>256</td>
<td>512</td>
</tr>
<tr>
<td>Total No. of incoherent integrations</td>
<td>5</td>
<td>1</td>
</tr>
<tr>
<td>Observation time</td>
<td>81.92 s</td>
<td>78.64 s</td>
</tr>
<tr>
<td>Spectral resolution</td>
<td>0.061 Hz</td>
<td>0.013 Hz</td>
</tr>
</tbody>
</table>

**Geostationary Operational Environmental Satellite-9 (GOES-9)** blackbody temperature (Tbb) data in the infrared (IR1: 10.2–11.2 μm) range, with horizontal 0.05° and 1-h resolutions, were downloaded from the Kochi University Web page (http://weather.is.kochi.u.ac.jp). The time series of Tbb at Kototabang is calculated by averaging four grid points around Kototabang.

The Tbb data and the temperature data in the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis, version 2, (hereinafter NCEP reanalysis; Kalnay et al. 1996) were used to estimate the cloud-top heights. From the NCEP reanalysis, the height distribution of temperature at 0°S, 100°E was computed by vertically interpolating temperature data and geopotential height data at standard pressure levels. Cloud-top height was defined to be the height at which the temperature was equal to Tbb. Because the time interval of the NCEP reanalysis data is 6 h, a linear time interpolation was applied to the temperature data to obtain hourly vertical profiles of temperature. Because cirroform cloud and nimbostratus are not regarded as perfect blackbodies, the estimated cloud top from an IR picture is usually lower than the actual cloud top (Sherwood et al. 2004). Because the current study only requires a level above which a cloud top is located, the estimated height is useful enough for discussion. Hereinafter, the estimated cloud top by this method will be referred to as the “Tbb cloud-top index.”

A weather radar operated at 9.74-GHz frequency installed at Kototabang (hereinafter referred to as the weather radar) was used to observe the vertical distribution of precipitation particles and to identify the stratiform rain region by detecting the bright band near the melting level. The weather radar was installed 15 m away from the EAR. The basic specification of this weather radar has been described in Konishi et al. (1998). Because of the maximum elevation angle of 29.5°, the vertical profile of the weather radar reflectivity factor (hereinafter referred to as $Z$) exactly above Kototabang was not observed. Therefore, the $Z$ obtained with the 29.5°-elevation beam is used to represent the values of $Z$ around Kototabang. Because the range resolution of the radar is 500 m, the vertical resolution is 246 m when the elevation angle is 29.5°. The interval of the scanning is around 10 min. After removing erroneous data, all of the 29.5°-elevation $Z$ data
observed in a circular direction around Kototabang were averaged to compute a representative \( Z \). By using the averaged \( Z \), the stratiform precipitation region was defined by judging the presence of a bright band. Because the bright band is at a height of about 4 km from the surface (865 m MSL), the horizontal distance of the bright band from the site in a 29.5°-elevation angle is only about 7 km. Therefore, the definition of a stratiform precipitation region using a bright band in this study is reasonable.

Because the horizontal distance from the radar exceeds 20 km for \( Z \) in the upper troposphere, the ratio of \( Z \) over noise level in a scanned circle was also calculated and is shown in the figures. If this value exceeds 50%, the possibility of the presence of \( Z \) over noise level just above the radar is thought to be high. The uniformity of cloud patterns around Kototabang was also checked by examining \( \text{GOES-9 IR pictures} \).

Outgoing longwave radiation (OLR) data were used to evaluate large-scale cumulus activity and to calculate any anomalies in cumulus activity from the long-year average around Indonesia.

3. Results

\( a. \) Selection of clusters

Because the climatological average of the rainfall in the area studied has a maximum in September–November (Hamada et al. 2002), November of 2003 is chosen for special observation of vertical motion in the cloud clusters. The monthly mean value of the cloud activity and wind distribution in November 2003 is close to normal, though the magnitude of the westerly flow in the lower troposphere and the easterly flow in the upper troposphere is slightly weaker than that in a normal year (not shown).

First, the cloud clusters that accompany active nimbostratus were selected. In this study, the cases in which upward motion was detected much more than downward motion in the middle and upper troposphere were selected with the method described below. In a height range of 8–12 km and for a continuous 3 h, the rate of \( W \) (150-m height and 12-min resolution) greater than 5 cm s\(^{-1}\) was calculated for the whole observational period, shifting the starting time by 1 h. If this rate was large, the upward motion was considered to be detected well.

The rate was calculated for 30 days in November of 2003, and three cases on 6, 8, and 20 November were selected as the cases in which upward motion was detected well. These three cases had a rate of larger than 70%. For these three cases, further examination was conducted with the following criteria: 1) a distinct bright band (local maximum of \( Z \) in the 4-6-km range) is detected over at least 3 h by weather radar (This is probably one of the most frequently used criteria for the definition of a stratiform precipitation region.), 2) the 3-h mean \( \text{GOES-9 Tbb} \) is less than 240 K, and 3) surface rainfall is observed. Criteria 2 and 3 indicate the presence of a well-developed precipitating cloud system. The three cases selected satisfy these criteria: the cases can be recognized as typical mesoscale convective systems, each with a well-developed stratiform precipitation region.

Here, the definition of a stratiform precipitation region is described. In basic terms, if a bright band is detected around the level of 0°C, the region is recognized as a stratiform precipitation region (Rosenfeld et al. 1995). This method leaves little possibility to select a convective region by mistake. Steiner et al. (1995) concluded that a considerable number of stratiform precipitation regions were detected without exhibiting a bright band, whereas only a few convective regions were observed with a bright band. This simple criterion was adopted because the intention of the current study is to select several typical stratiform precipitation periods rather than to apply a strict classification between convective and stratiform precipitation regions. The validity of the classification of stratiform precipitation regions was examined further by uniformity of Tbb values around Kototabang.

Although it is difficult to identify the exact extent of nimbostratus with our dataset, clouds embedded in the middle and upper troposphere of a stratiform precipitation region are classified as the nimbostratus in the following description, because such clouds have many properties of nimbostratus (see Houze 1993, chapter 1.2).

\( b. \) Case of 6 November

Figure 1 shows the time sequence of a mesoscale cloud cluster passing over Kototabang in the evening of 6 November. The local time in Sumatra is +7 h from UTC time. It passes over Kototabang from the southeast to the northwest. The Tbb at Kototabang reaches a minimum value of less than 200 K around 1245 UTC. The region of the lowest Tbb values moves to the northwest of Kototabang by 1645 UTC while the relatively low Tbb values of about 230 K still surround the Kototabang region.

Figure 2 shows the time series of \( W \), the Tbb cloud top index, \( Z \), and surface rain at Kototabang. Until a bright band appears at around 5 km at 1340 UTC (Fig. 2b), convective activity appears to pass over the Kototabang region, because the Tbb cloud-top index reaches 15 km (1245 UTC) and \( W \) has a magnitude greater than...
After the passage of the possible convective activity, a bright band appears after 1340 UTC and continues until 1900 UTC, although gradually becoming weak in the later stage. This period (1340–1900 UTC) is regarded as a stratiform precipitation region. The Tbb cloud-top index is near 14 km at 1345 UTC and gradually decreases to 12 km at 1645 UTC. The spatial distribution of Tbb is mostly uniform around Kototabang (Fig. 1). Surface rain is observed during almost all of the stratiform precipitation period (until 1820 UTC), and the rate of rainfall is less than 4 mm h$^{-1}$. This weak rate of surface rain supports the fact that this period falls under stratiform precipitation conditions (Balsley et al. 1988).

A noticeable feature in the nimbostratus is the continuous gentle upward motion with small fluctuations in the later stage of the stratiform precipitation region. In examining the stratiform precipitation region in detail, a slight increase in $Z$ and the surface rain can be detected around 1600 UTC, as well as a strong upward motion of about 1 m s$^{-1}$ and downward motions in the 6–12-km height range. At this time, convective clouds could possibly be embedded in the stratiform precipitation region. After this event (around 1600 UTC), gentle upward motions with small fluctuations are continuously observed in the middle and upper troposphere.

Around 1630 UTC, the layer of gentle upward motion, where the value of $W$ is positive and smaller than 40 cm s$^{-1}$ over most of the range, has a large vertical extent (6.0–11.0-km height range). Figure 3b shows the temporal distribution of the fine structure of $W$ in the
FIG. 2. Vertical motion and cloud variables when a cloud cluster passes over Kototabang on 6 Nov 2003. (a) Shading and contours show 12-min-averaged vertical motion (cm s$^{-1}$) observed by the EAR. White area indicates missing data. Thick contour levels are $-300, -200, -100, -40, 40, 100, 200, and 300$ cm s$^{-1}$; thin contour levels are from $80$ to $+80$ cm s$^{-1}$ with a $20$ cm s$^{-1}$ interval except the level of $0$ cm s$^{-1}$. The values of Tbb cloud-top index inferred from GOES-9 Tbb data are shown by the open circles and thick line (see section 2 for detail). Altitude of $0$ km is MSL. (b) Radar reflectivity factor $Z$ derived from the weather radar around Kototabang. Shading and solid contour lines show the averaged $Z$ around EAR (dBZ). The contour interval is $5$ dBZ. The broken lines indicate the rate (%) of returned reflectivity in a scanned circle (see section 2 for detail). (c) Ten-minute accumulated rain amounts by an optical rain gauge (mm h$^{-1}$).
Fig. 3. Fine structure of the vertical motion (cm s$^{-1}$) when a cloud cluster passes over Kototabang on 6 Nov 2003. 
(a) Vertical motion [shown in (b)] averaged in 8–11-km height range. (b) Twelve-minute average of vertical motion. White areas indicate missing data. Thin contours are for each 10 cm s$^{-1}$ between $-100$ and $100$ cm s$^{-1}$, and thick contours are for each 50 cm s$^{-1}$ between $-300$ and $300$ cm s$^{-1}$. (c) Three-minute average of vertical motion. Contours and shading patterns are same as in (b).
nimbostratus. The vertical range of the layer seems to be lifted to 8.0–12.5 km around 1830 UTC. During 1630–1830 UTC, upward motions of greater than 40 cm s\(^{-1}\) and downward motions are almost absent in the height range. After 1830 UTC, the range of weak upward motion is dissolving.

The most striking feature among the various structures of \(W\) in the nimbostratus is well shown around 8–11 km during 1630–1830 UTC. Only a few contours are drawn at the 10 cm s\(^{-1}\) interval in Fig. 3b; both the time and height variations of \(W\) are inactive over this period. Figure 3a exhibits the vertically averaged \(W\) in the nimbostratus (8–11 km). The vertically averaged \(W\) is mostly constant in most of the stratiform precipitation period; it is in the 15–40 cm s\(^{-1}\) range between 1430 and 1800 UTC except for a large fluctuation around 1530 UTC. Another dataset with a finer time resolution of 3 min was inspected (Fig. 3c), though the data are missing for some regions. Significant fluctuations of \(W\) in the dataset with a time scale of less than 12 min were not detected.

Figure 4a shows the hourly histogram of \(W\) in height range of 8–11 km on 6 November with 12-min and 150-m-height resolutions. The number of values of \(W\) was counted during 1 h. Prior to 1400 UTC, the frequency of downward motion is almost the same as that of upward motion, which suggests a convective feature. After 1400 UTC, the frequency of stronger upward motion of greater than 40 cm s\(^{-1}\) slightly increases, whereas that of downward motion becomes less than 20%. During the later stage of the stratiform precipitation region (1700–1800 UTC), the frequency of weak upward motion (0–40 cm s\(^{-1}\)) exceeds 95%. Most of the weak upward motions during the 1700–1800 UTC period are of a gentle nature in nimbostratus.

c. Case of 8 November

The time series of the cloud clusters on 8 November is shown in Fig. 5. A cluster passes over Kototabang from east to west. The movement is traced well with a 200-K Tbb contour from 1045 to 1245 UTC. At 1045 UTC, the region with Tbb of lower than 200 K is detected east of Kototabang (between 0.5°N, 100.7°E and 0.7°S, 101.2°E). At 1245 UTC, one of the clusters is located north of Kototabang (0.25°N, 100.25°E). After 1445 UTC it moves to the west and the shape of the low Tbb region (less than 200 K) becomes unclear. The wide region with the value of moderately low Tbb (200–230 K) extends around Kototabang.

Figure 6 shows the time series of \(W\), the Tbb cloud-top index, \(Z\), and surface rain. The surface rain starts just before 1200 UTC. Strong upward motions of about 1 m s\(^{-1}\) are sporadically detected during 1200–1400 UTC between 6 and 13 km. A bright band is not detected prior to 1230 UTC. Although the rate of surface rain is not very high, that is, 5–8 mm h\(^{-1}\), the strong magnitude of \(W\) and the lack of a bright band suggest that the period between 1200 and 1230 UTC is possibly defined as a convective precipitation region.

Between 1230 and 1700 UTC, a bright band in \(Z\) and nonzero surface rain are almost continuously observed. The Tbb cloud-top index is 14 km at 1345 UTC and gradually lowers to 12 km around 1645 UTC. Low Tbb values of less than 230 K extend around Kototabang (Fig. 5). These results indicate that Kototabang is under
the conditions of a stratiform precipitation region for the period of 1230–1700 UTC.

Weak upward motion with small fluctuations is observed in the later stage of the stratiform precipitation (1400–1700 UTC) between heights of 7 and 11 km, as was found in the case of 6 November. Over most of the region, the magnitude of the upward motion is less than 40 cm s$^{-1}$, although stronger $W$ is found in the 8–10-km range around 1500 UTC. The maximum dominance of gentle upward motion is confirmed in the 8–11-km range at 1600–1700 UTC in Fig. 4b, when the ratio of gentle upward motion is nearly 100%.

Figure 7 exhibits the detail inside the region of gentle upward motion in two datasets with different time resolutions. Between 1530 and 1700 UTC, the average of $W$ in the height range of 8–11 km is 10–30 cm s$^{-1}$ (Fig. 7a). At that period, the gentle upward motion is also described well with 3-min-resolution data (Fig. 7c). Most of the notable features in Fig. 7c can be seen also in Fig. 7b, at least in the region of gentle upward motion detected in Fig. 7b. Vertical motion features with a period of shorter than 12 min are not predominant in this period.

The 6 and 8 November cases are similar in the context of diurnal change. Both cases seem to have the same features, such as turbulent convective mixing in the afternoon (before 1300 UTC, or 2000 LST) and stratiform precipitation in the evening (after 1300 UTC). Previous studies also have shown a similar feature in diurnal change of convection over Sumatra (Mori et al. 2004; Renggono et al. 2001).

d. Case of 20 November

Figure 8 shows the time series of a cloud cluster on 20 November. This cluster rapidly develops around 1345 UTC around Kototabang. The low Tbb region (less than 230 K) around Kototabang extends to the west, gradually expanding, perhaps due to strong upper tropospheric easterly flow.
The cluster has a very high Tbb cloud-top index around 1345–1445 UTC (Fig. 9a). The index reaches 15.5 km and the radar echo top (Fig. 9b) is higher than 13 km around 1440 UTC. A bright band, which appears at 1430 UTC, continues until 2200 UTC. The rainfall is observed over almost the entire period, although the rate of the rain is much less than that in the previous two cases (continuously less than 1.2 mm h⁻¹). Spatially uniform values of Tbb are produced around Kototabang after 1445 UTC, as found at 1545, 1745, and 1945
Fig. 7. The same as Fig. 3, but for the 8 Nov cloud cluster case.
UTC (Fig. 8). From the above-mentioned results, the period of 1430–2200 UTC is regarded as a stratiform precipitation region.

Gentle upward motion ($0$–$40$ cm s$^{-1}$) with small fluctuations is detected in the later stage of the stratiform precipitation (1900–2100 UTC). The height range of the region is 6–11 km at around 1900 UTC, and the lower edge of the region gradually lifted to 7.5 km at 2100 UTC. Upward motion of $40$–$60$ cm s$^{-1}$, however, is found around 9 km during 1900–2010 UTC. When considering that values of $Z$ greater than 15 dBZ extend above 8 km around 1900 UTC, a relatively active part of the nimbostratus exists around Kototabang. The dominance of gentle upward motion over heights of 8–11 km can be confirmed in Fig. 4c. In this case, the frequency of weak upward motion ($0$–$40$ cm s$^{-1}$) was the highest during 2000–2100 UTC, though the rate is somewhat less than that in the previous two cases. The predominant feature in 3-min data (not shown) is also captured well in 12-min data in Fig. 9a.

e. Some comments on the case studies of 6, 8, and 20 November

Gentle upward motion with small fluctuations in the later stage of the stratiform precipitation region is commonly observed in these three cases, except for the small differences in the magnitude of $W$.

One interesting case study with VHF radar has been obtained by Cifelli and Rutledge (1994). In the study, typical cloud clusters were selected for the monsoon and break seasons in northern Australia in which the evolutions of the stratiform precipitation region were documented. In a break-season case, continuous upward motions with magnitudes of less than 50 cm s$^{-1}$ were observed (Fig. 7 in Cifelli and Rutledge 1994). Although the current results are similar to those of the previous study, the vertical extension and length of pe-

![Fig. 8. The same as Fig. 1, but for the 20 Nov cloud cluster case at (a) 1345, (b) 1545, (c) 1745, and (d) 1945 UTC.](Image)
The time resolution of 3 min is sufficiently smaller than the scale of the buoyancy period (7–10 min): the lower limit of internal gravity wave. If 10–15 m s⁻¹ is adopted as a typical moving speed of a cloud system, the horizontal resolutions would be 1.8–2.7 km in 3-min
data. It is probably smaller than the scale of each cumulonimbus cell.

4. Discussion of the mechanism that produces gentle upward motion in nimbostratus

Here the discussion is centered on the mechanism for producing the observed gentle upward motion (0–40 cm s⁻¹) with small fluctuations in nimbostratus during the later stage of the stratiform precipitation region in the three mesoscale cloud clusters.

To keep a gentle upward motion with small fluctuation for a long period, needed are two conditions: 1) the dominance of processes to maintain slow upward motion in a wide region and 2) the suppression of processes that make short-period and/or strong vertical motion. Even within the stratiform precipitation regions, gentle upward motions are not always detected. Instead, the dominance of short-period variability appears, like the early stage of the stratiform period in the 6 November case (Fig. 3b), or downward motion is widely observed inside the clouds, like the early stage (around 1700 UTC) in the 20 November case (Figs. 4c and 9b). Hereinafter, we discuss what physical processes should be dominant or suppressed to form gentle upward motion with small fluctuations.

The following processes can produce a uniform gentle upward motion. Latent heat release resulting from deposition growth is one of the most plausible causes for the uniform upward motion. If the air around ice particles is supersaturated relative to ice, much vapor can be deposited on the ice particles. Latent heat release by deposition can occur over a large part of the height range and enables development of a widely extended mesoscale low pressure region in the nimbostratus (e.g., Houze 1989).

Gravity waves with time and spatial scales comparable to or larger than the extension of the nimbostratus region are another candidate. They are induced by continuous heating produced by the total effects of the cumulonimbi that occur repeatedly and frequently in the convective region. When the horizontal and vertical scales of gravity waves are large enough, one large nimbostratus region can be in one upward phase region of the wave. Pandya and Durran (1996) demonstrated that divergent flow with slowly rising air in the upper troposphere in the stratiform cloud region could be interpreted as a phase of the gravity wave induced by latent heat release. However, note that a pair of cell-type upward and downward motions was embedded in the positive W phase region of the gravity wave response in the stratiform region (e.g., see Fig. 11 in Pandya and Durran 1996). Further study is required to determine if such gravity waves can actually produce gentle upward motion.

On the other hand, the following processes tend to make fluctuations with short time scales. One process is old cells, which have their origin in the convective region and are often observed even far from the convective part in a strong vertical shear condition (e.g., Houze 1993) in the upper part of nimbostratus. If these cells have enough buoyancy, they make a strong upward motion with small scale. Another process is the seeder–feeder mechanism that sometimes produces a rainband structure. Latent heat is released by the freezing of supercooled raindrops when falling ice particles come in contact with the drops (e.g., Hobbs et al. 1980). This may happen in the lower part of nimbostratus and may induce thermal instability, resulting in a small-scale cell-like structure. Gravity waves with smaller time scales than the duration of the nimbostratus, which are produced possibly by each convective cloud and propagate to the stratiform part, also create large variability. These processes should be suppressed when gentle upward motions without small-scale fluctuation dominate in the stratiform precipitation regions.

In this study, it is not clear which processes are active and/or which are suppressed in the nimbostratus because of a lack of information on the cloud microphysics and horizontal distribution of the cloud systems. To examine the deposition and liming processes, instruments that can obtain microphysical data, such as millimeter cloud radar and cloud vide sondes, are necessary. For evaluation of the gravity-wave sources, a weather radar that can scan more widely around the VHF radar would be more useful than that used in this study, which has a limited scan region because it is installed very close to the EAR and has only a low-elevation angle (up to 29.5°).

Also important are numerical studies with a model with details of ice processes and gravity waves with various scales. In those studies, the knowledge of the precise distribution of upward motion should become a clue for appropriate simulation. Our results on the uniform upward motion with small fluctuation in the later stage of the stratiform precipitation region can be available when evaluating the model results quantitatively, because they give accurate magnitude of variability in vertical motion.

5. Conclusions

The fine structures of vertical motion W in three mesoscale cloud clusters were investigated by the EAR. By applying an observational mode to observe intensively in the vertical direction, W was observed with
sufficient accuracy (better than 5 cm s⁻¹) and fine resolutions (3 and 12 min; 150 m in the vertical direction) to examine weak vertical motions in nimbostratus.

The passage of several typical isolated mesoscale cloud clusters was observed. In the later 2–3 h of the stratiform precipitation period, the gentle upward motion with small time and height fluctuations was observed in three cases (6, 8, and 20 November). Over approximately 3–5 km from the middle to upper troposphere, W had weak positive values (0–40 cm s⁻¹) almost continuously, with hardly any upward motions greater than 40 cm s⁻¹ or any downward motion. Within these regions, the time and height fluctuations of W were small. The uniqueness of this paper is in the quantitative description of the gentle upward motion with small fluctuation. The important point is not only that the averaged value of W was observed to be on the order of 10 cm s⁻¹, as is well known, but also that the uniform weak upward motion was found to continue in a deep height range over 2–3 h in the stratiform precipitation region.

An observation of VHF radar is limited to only above the observational site. To obtain the complete spatial structure of precipitation clouds around Kototabang, a Doppler weather radar was operated at about 20 km southeast of Kototabang in October and December of 2005. Further studies with this Doppler weather radar and the EAR will be made to investigate relationships between W and cloud distribution.

Because this study addresses only three cases in November of 2003, the statistical frequency of the occurrence of gentle upward motion cannot be obtained. It is of interest to examine how often a gentle upward motion with small fluctuation is seen in mesoscale cloud systems and what factor in the basic state controls the magnitude and variability of upward motion in a nimbostratus region. An extended study with longer routine records is in progress to reveal these features.

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