The Generation of Turbulence below Midlevel Cloud Bases: The Effect of Cooling due to Sublimation of Snow

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ABSTRACT

In the author’s experience as a forecaster, commercial aircraft sometimes report turbulence beneath midlevel clouds that extend above upper frontal zones. Turbulence caused by Kelvin–Helmholtz instability occurs in upper frontal zones with strong vertical shear of horizontal winds. However, the turbulence seems to occur not only in the cloud bases (where upper frontal zones are) but also below the cloud bases where the vertical shear is not strong. Because those clouds are usually accompanied by precipitation that does not reach the ground, cooling by evaporation or sublimation seems to contribute to the generation of turbulence. In this paper, the mechanisms generating turbulence below midlevel cloud bases are examined by using observations and high-resolution three-dimensional numerical simulations with idealized initial conditions. The numerical simulations showed that the following sequence of events led to turbulence. Falling snow sublimated below cloud bases and cooled the air, which created absolute instability. This generated Rayleigh–Bénard convection cells. The vertical motion caused turbulence. The horizontal scale of the convection was about 800–1000 m, and the variations of vertical wind velocity were up to about 7 m s\(^{-1}\). The cloud base was accompanied by a virga-like distribution of snow. Sensitivity experiments showed the appropriate conditions to cause the turbulence: 1) the cloud-base temperature was between about 0°C and −15°C, 2) the relative humidity in subcloud layers was sufficiently low, and 3) the stability in subcloud layers was weak. The results of the numerical simulations agreed well with the observations.

1. Introduction

In aircraft, a sudden encounter with strong turbulence is potentially dangerous for both the passengers and crew. Even when the turbulence is not strong, passengers can feel uncomfortable. For both the safety and comfort of flight, accurate predictions of turbulence are required in order to avoid the zones where it occurs.

To achieve better predictions, the generation mechanisms of turbulence need to be understood. Kelvin–Helmholtz (KH) instabilities (Browning et al. 1973; Lilly 1986; Fritts et al. 2003), mountain waves (Scorer 1949; Durran 1986; Cohn et al. 2011), and convective clouds (MacCready 1964; MacPherson and Isaac 1977) are widely known to generate turbulence encountered in flight. Other causes have been recently studied. Luce et al. (2009) observed vertical wind oscillations of up to ±1.5 m s\(^{-1}\) beneath cirrus clouds by using the middle and upper atmosphere radar, a Rayleigh–Mie–Raman lidar, and a balloon radiosonde; according to them, sublimation of ice crystals in precipitation falling beneath a cloud, or cloud-base detrainment instability (Emanuel 1981), may have caused convective instability (rather than dynamical shear instability) that created the vertical wind oscillations. Using high-resolution numerical simulations, Lane et al. (2003) and Lane and Sharman (2008) investigated turbulence caused by the breaking of convectively induced internal gravity waves or KH waves, above and near cumulonimbus clouds. According to Knox (1997), McCann (2001), and Koch et al. (2005), internal gravity waves produced by highly unbalanced flows can cause turbulence through the local modification of the environmental Richardson number to be less than the critical value of 0.25.

In the author’s experience as a forecaster, commercial aircraft sometimes report turbulence beneath midlevel clouds that extend above upper frontal zones; those clouds are usually accompanied by precipitation that does not reach the ground, also known as virga. It is well known that turbulence occurs in upper frontal...
zones by KH instability with strong vertical shear of the horizontal winds. However, turbulence below midlevel clouds seems to occur not only in the upper frontal zones but also below the frontal zones where vertical shear is not strong. Sometimes the turbulence occurs several thousand feet (1 ft ≈ 0.305 m) below the frontal zones. This is a very significant feature of the turbulence below the bases of midlevel clouds. The turbulence seems to occur more frequently below midlevel clouds than below high- or low-level clouds. From this point forward, we refer to this turbulence as midlevel cloud-base turbulence (MCT). Most commercial airlines are affected by MCT because the bases of midlevel clouds lie in the same altitude range in which jet aircraft ascend and descend and commuter aircraft cruise. Therefore, it is important for aviation communities to understand the generation mechanism and occurrence conditions of MCT.

Since MCT usually occurs below the bases of midlevel clouds accompanied by precipitation that does not reach the ground, cooling by evaporation or sublimation seems to contribute to the turbulence. Although some studies have addressed the effects of cooling below cloud bases, few studies have examined the relation between cooling and turbulence. Harris (1977) performed one-dimensional numerical calculations in which the pressure and temperature at cloud base were fixed at 650 hPa and 261 K, respectively, and showed that sublimation of solid precipitation caused cooling that led to an absolutely unstable layer. The lapse rate below the cloud base was higher with a greater amount of solid precipitation and with decreasing humidity in the subcloud layer. Using a vertically pointing Doppler radar, he also observed holes and stalactites (appendages extending downward) in the base of a uniformly generated layer of precipitation. He deduced that updrafts and downdrafts of 1–3 m s⁻¹ correspond to the holes and stalactites, and that they had scales of 500 m in the vertical and 500–1500 m in the horizontal. However, he did not study turbulence produced after the absolutely unstable layer formed, and he based his calculation on one-dimensional equations that could not represent three-dimensional motions.

Kanak and Straka (2006) and Kanak et al. (2008) performed high-resolution three-dimensional numerical simulations of mammatus on a portion of a cumulonimbus cirrus anvil. The initial conditions were derived from observed thermodynamic soundings that featured a dry-adiabatic subcloud layer with low relative humidity. They showed that mammatus clouds formed in their simulations and tested formation mechanisms for mammatus; they concluded that the cooling due to sublimation was the largest term producing negative buoyancy in the mammatus.

The environmental conditions of dry microbursts (or low-reflectivity microbursts) look similar to those of MCT. Wakimoto (1985) defined dry microbursts as “a microburst that is accompanied by little or no rain between the onset and the end of the high winds, including intermediate calm periods, if any. This type of microburst is usually associated with virga from altocumuli or shallow, high-based cumulonimbi.” He proposed and illustrated a conceptual model of the environment favorable for dry microbursts: a deep, dry-adiabatic, and very dry subcloud layer from the surface to a midlevel cloud. By using radar observations of the bright band combined with numerical simulations with a one-dimensional microphysical model, Wakimoto et al. (1994) concluded that the sublimation of snowflake aggregates is the primary mechanism forcing dry microbursts. They also found that virga associated with the downdrafts consisted of ice particles rather than raindrops because the radar bright band was located at the visible termination of the virga (i.e., the melting level).

The purpose of this study is to examine the generation mechanism and occurrence conditions of MCT by using upper-air soundings, aircraft observations, and high-resolution three-dimensional numerical simulations with idealized initial conditions. Section 2 of this paper presents observations for a typical case of MCT. Section 3 gives results of numerical simulations with idealized initial conditions, and of sensitivity experiments for the occurrence of MCT. We compared these results with the observations. The similarity and differences between MCT and mammatus clouds and those between MCT and dry microbursts are discussed in section 4. A summary is presented in the final section.

2. Observations of a typical case of midlevel cloud-base turbulence

a. Overview

Around 0600 UTC 12 May 2008, many aircraft encountered turbulence over the Kanto district in central Japan (Fig. 1). The turbulence was observed between flight levels from FL080 to FL270 (≈2400–8200 m) over the Kanto district (Fig. 2). (“FL” indicates altitudes in hundreds of feet in the International Standard Atmosphere. Hereinafter, FL and “ft” indicate altitudes in the International Standard Atmosphere and “meter” and “m” indicate altitudes above mean sea level.) At this time, there was a typhoon south of Japan (Fig. 2a) that was moving to the northeast. Observations suggest that the turbulence occurred in convective clouds associated with the typhoon; however, radar reflectivity in the south part of the Kanto district was not strong (Fig. 3a); that is, the turbulence did not occur in the convective clouds. Precipitation was not observed at Tokyo International Airport (HND in
Fig. 1b) until 1330 UTC. At this time, the midlevel atmosphere was very dry, and moist air was being advected above the dry air over the Kanto district (Fig. 4a).

b. Observations

1) RADAR REFLECTIVITY

In horizontal and vertical depictions of radar reflectivity through the turbulent region (Fig. 3), somewhat stronger radar reflectivity (>10 dBZ), slanting toward the north, was found at heights from about 2500 to 9000 m to the north of 35°N (Fig. 3b) whereas reflectivity was very weak at a level of 2000 m over the Kanto district (Fig. 3a). That is, there was precipitation in the midlevels of the atmosphere but it did not reach the ground. Turbulence was reported between FL080 and FL180 (≈2400 and ≈5500 m) around 35°N (red dashed rectangle in Fig. 3b), indicating that the
turbulence occurred in the lower part of the volume of radar reflectivity.

2) TIME SERIES OF PILOT REPORTS AND AUTOMATIC AIRCRAFT OBSERVATIONS

A front with wind shear and a small temperature gradient was found at 20 000 ft (∼6100 m) at 0000 UTC in the time series depiction of pilot reports (PIREP) and Aircraft Meteorological Data Relay (AMDA) reports (Fig. 5). It lowered with time and reached 13 000 ft (∼4000 m) at 0900 UTC (thick solid gray line in Fig. 5). Clouds with somewhat stronger radar reflectivity in Fig. 3 probably formed above the front. As shown in Fig. 5, turbulence occurred not only at the front but also more than 5000 ft (∼1500 m) below the front (especially from 0000 to 0730 UTC).

3) UPPER-AIR SOUNDING

In the upper-air sounding at 1200 UTC 12 May 2008 at Tateno, the nearest radiosonde station to the “TATEYAMA” reporting point, a sharp temperature inversion (which indicated an upper frontal zone) with wind shear can be seen at a level of about 500 hPa, with saturated moist air above it and dry air beneath it (Fig. 4a). The stability was nearly dry adiabatic from the bottom of this inversion to about 600 hPa. Turbulence was reported between approximately 500 and 600 hPa around Tateno at this time (data not shown); that is, turbulence occurred not only at the front but also in the nearly dry-adiabatic layer below the front. Vertical profiles of wind speed, direction, and vertical shear (Fig. 4b) from this sounding indicate that winds changed direction near the front and at the bottom of the nearly dry-adiabatic layer (at about 620 hPa), while such changes were small in the middle of the nearly dry-adiabatic layer above 620 hPa. Although turbulence in the upper frontal zone could have been caused by KH instability with strong vertical shear, the turbulence in the nearly dry-adiabatic layer was probably caused by other mechanisms.

4) AIRCRAFT OBSERVATIONS

A jet aircraft encountered moderate turbulence while descending to HND via the “XAC” and TATEYAMA reporting points (positions in Fig. 1b) around 0615 UTC 12 May 2008. The vertical acceleration (Fig. 6a) varied from 0.56 to 1.43 times the gravitational acceleration from altitudes of 16 500 to 8500 ft (∼5000–2600 m), indicating that the airliner encountered turbulence between these altitudes. There was a temperature inversion around 14 000 ft (∼4300 m) (Fig. 6b), which corresponded to the upper frontal zone in Figs. 4a and 5. Between 10 000 and 13 000 ft (∼3000 and 4000 m), the potential temperature (Fig. 6c) decreased with height, and the lapse rate (Fig. 6d) exceeded the dry-adiabatic
lapse rate ($9.76 \text{ K km}^{-1}$), so that the atmosphere was absolutely unstable below the inversion. Turbulence was continuously observed between the inversion and the absolutely unstable layer. Wind speed (Fig. 6e) decreased rapidly with height at the bottom of the inversion layer, and became very slight in the absolutely unstable layer. Vertical shear was relatively strong in the inversion and at the bottom of the absolutely unstable layer (Fig. 6g), because of the vertical changes in wind speed and direction (Fig. 6f). The Richardson number was approximately 0.25 (the critical value for turbulence generation) in the inversion layer, while it was negative in the absolutely unstable layer (Fig. 6h), suggesting that the turbulence was caused by KH instability in the inversion layer and by absolute instability below the inversion.

c. Summary

Around 0600 UTC 12 May 2008, many aircraft encountered moderate turbulence in the midlevels of the atmosphere. The turbulence occurred not only at the front but also more than 5000 ft ($\approx 1500 \text{ m}$) below the front. The air was dry below the front while it was moist above the front. Somewhat stronger radar reflectivity was observed above and around the front but it did not reach the ground. The stability and vertical shear were strong in the frontal zone while the stability was nearly dry adiabatic and the vertical shear was weak in the middle of the nearly dry-adiabatic layer.

3. Numerical simulations

a. Numerical model

To investigate the generation mechanism and occurrence conditions of MCT, we performed high-resolution three-dimensional numerical simulations with a mesoscale model of the Japan Meteorological Agency (Saito et al. 2007). The operational mesoscale model has a fully compressible nonhydrostatic dynamical framework and a horizontal grid spacing of 5 km. The precipitation process is treated with a combination of an explicit three-ice bulk microphysics scheme (Ikawa and Saito 1991) and a Kain–Fritsch scheme (Kain 2004). The model forecasts six mixing ratios (for water vapor, cloud water, cloud ice, rain, snow, and graupel) and one number concentration (for cloud ice). The model assumes spherical particles with constant density for all hydrometeors. The size distribution for rain, snow, and
graupel are assumed to follow an exponential function and those for cloud water and cloud ice are assumed to be monodispersed. Cloud microphysical processes are treated directly depending on the size distributions except for some basic processes (e.g., nucleation of cloud particles, conversion from cloud particles to precipitation particles), which are treated by parameterizing the processes. The fall velocities of the precipitable hydrometeors (cloud ice, rain, snow, and graupel) are given by the mean fall velocity weighted by the mass distribution.

In this study, we ran the mesoscale model with a grid spacing of 50 m in all directions and a time step of 0.1 s. The horizontal domain was 4000 m × 4000 m (80 × 80 grids) with a periodic boundary condition, and the vertical domain extended from 0 to 10 000 m (201 grids). With this grid spacing and domain, it is almost possible to resolve phenomena with a horizontal scale from about 100 to 2000 m, which induce the strongest turbulent response from large commercial aircraft (Lane et al. 2012). The model used a Deardorff scheme (Deardorff 1980) for subgrid turbulence, and did not take into account the surface flux since it is likely irrelevant to MCT. The other configurations were the same as in the operational model.

b. Initial conditions

As shown above, MCT seems to occur below midlevel clouds accompanied by precipitation that does not reach the ground. We set up idealized initial conditions to better depict the situation. At first, the initial conditions...
were set to be horizontally homogeneous; however, in order to introduce fluctuations, we then added random perturbations on the scale of 1% to the wind components and the relative humidity.

The vertical profiles of horizontally averaged temperature and dewpoint temperature at the initial time (Fig. 7a) consist of three parts: a moist region (above 4500 m), a stable layer (from 4000 to 4500 m), and a dry...
region (beneath 4000 m). These regions correspond to a cloud layer, a cloud base, and a subcloud layer, respectively. The relative humidity was set to 100% in the moist region and 40% in the dry region, and was linearly interpolated in the stable layer. The increasing rate of equivalent potential temperature was set to 3 K km\(^{-1}\) in the moist region, and the lapse rate of temperature was set to 8 K km\(^{-1}\) in the dry region. A uniform temperature of \(-10^\circ\)C was assumed in the stable layer. Also in the initial conditions, the horizontal wind speed was set to increase with height, to keep the Richardson number \(R_i = 1.5\) in the stable layer and \(R_i > 1.5\) everywhere else (Fig. 7b). The vertical wind speed component was set to zero, and no hydrometeors were assumed to exist at any level (data not shown). The results of the control experiment are given in section 3c, and those of the sensitivity experiments are given in section 3d.

c. Results of the control experiment

Horizontal and vertical sections of the control experiment at times \(t = 77, 104,\) and 125 min are shown in Fig. 8. The hydrometeors were generated in the moist region in the early phase of integration (before \(t = 77\) min). Though the model forecasts five types of hydrometeors (cloud water, cloud ice, rain, snow, and graupel), almost all of the generated hydrometeors were snow (data not shown). The snow then gradually descended with time and penetrated below 4500 m (Fig. 8d). Snow sublimated there, which cooled the atmosphere and generated an absolutely unstable layer (the lapse rate \(> 9.76\) K km\(^{-1}\)) below the stable layer. The depth of the unstable layer expanded with time and reached about 450 m (\(z = 3550–4000\) m) at \(t = 77\) min (Figs. 8j and 9b). Observable changes in the vertical winds cannot be seen until \(t = 77\) min (Fig. 8g). After this time, convection cells with
orderly up- and downdrafts and a virga-like distribution of snow appeared beneath the stable layer, and the vertical winds attained a maximum strength at \( t = 104 \) min (Figs. 8b, 8e, and 8h). The horizontal wavenumber perpendicular to the flow was about 4–5; that is, the horizontal wavelength was about 800–1000 m. The convective cells disappeared quickly after the convection peaked and can no longer be seen at \( t = 125 \) min (Figs. 8c, 8f, and 8i). The occurrence of convection cells under absolutely unstable conditions suggests that the cells are Rayleigh–Bénard convection (see discussion at the end of this section).

Horizontally averaged vertical profiles in the control experiment are shown in Fig. 10. In the early integration phase of the control experiment, snow was generated in the moist region and most of the snow later sublimated in a layer several hundred meters below the stable layer (Fig. 10a). No rain was produced at any level, and the amounts of graupel and cloud ice were negligible (data not shown). The snow mixing ratio gradually decreased with time because water vapor was not resupplied in the present simulation. The accumulated amount of sublimation of snow mixing ratio since the initial time \( dR_s \) was greatest at the bottom of the stable layer (at \( z = 4000 \) m) at \( t = 77 \) min (\( dR_s = -0.383 \) g kg\(^{-1}\)), and later at \( t = 125 \) min (\( dR_s = -0.643 \) g kg\(^{-1}\)) (Fig. 10b). Temperature decreased below the stable layer at \( t = 77 \) min (Fig. 10d). The temperature decrease from the initial time (\( \delta T \)) at \( t = 77 \) min was largest at the bottom of the stable layer (Fig. 10c). This decrease (\( \delta T = -1.16 \) K) was in close agreement with \( L_s \delta R_s/C_p = -1.08 \) K, where \( L_s \) is the latent heat of the sublimation of snow or ice (\( L_s = 2.834 \times 10^6 \) J kg\(^{-1}\)) and \( C_p \) is the specific heat at constant pressure (\( C_p = 1005 \) J kg\(^{-1}\) K\(^{-1}\)). Thus, most of the temperature decrease resulted from the sublimation of snow. At \( t = 77 \) min, the lapse rate increased to about 10.4 K km\(^{-1}\) below the stable layer (at \( z = 3750 \) m) and decreased to about \(-4.12 \) K km\(^{-1}\) in the stable layer (at \( z = 4400 \) m) (Fig. 10f). Potential temperature decreased with height from 3550 to 4000 m (Figs. 9b and 10c). That is, the absolutely unstable layer appeared below the cloud base while stability increased above the unstable layer. These results are consistent with the upper-air sounding and aircraft observations, which show a sharp inversion and a subsequent nearly dry-adiabatic or absolutely unstable layer (Figs. 4a and 6). At \( t = 125 \) min, the absolutely unstable layer disappeared and atmospheric conditions became nearly dry adiabatic at about 1000 m below the bottom of the stable layer (Figs. 10c and 10f). Vertical wind speed remained at almost zero until \( t = 77 \) min and, then, increased to a maximum at \( t = 104 \) min. The vertical wind speeds ranged from about \(-3.8 \) to \(+3.1 \) m s\(^{-1}\) (i.e., the difference was about 6.9 m s\(^{-1}\)). According to WMO (2003), the quantitative severity of turbulence can be related approximately to the derived equivalent vertical gust velocity. The gust velocities of 2–4.5, 4.5–9, and >9 m s\(^{-1}\) correspond to light, heavy, and severe turbulence, respectively; therefore, the orderly up- and downdrafts in this simulation can cause heavy turbulence. Though the orderly up- and downdrafts rapidly decayed after \( t = 104 \) min, the vertical wind speeds ranged about \( \pm 1.6 \) m s\(^{-1}\) (the difference was about 3.2 m s\(^{-1}\)) at \( t = 125 \) min; this corresponds to light turbulence. These results are consistent with the observations, which show the existence of turbulence below midlevel cloud bases (Figs. 5 and 6). There was almost no change in the horizontal wind speeds until \( t = 77 \) min; however, at \( t = 125 \) min, when the convective mixing progressed, a layer with a homogeneous wind speed of about 12 m s\(^{-1}\) formed in the middle of the nearly dry-adiabatic layer (\( z = 3200–4000 \) m) (Fig. 10h). As a consequence, vertical shear decreased there and increased at the top and

**Fig. 9.** Vertical profiles for horizontally averaged potential temperature in the control experiment at \( t = (a) 55, \) (b) 77, and (c) 104 min.
bottom of the nearly dry-adiabatic layer (Fig. 10i), due to the vertical redistribution of momentum. These results also are consistent with the upper-air sounding and aircraft observations, which show relatively strong vertical shear in the inversion and at the bottom of the nearly dry-adiabatic layer, and weak vertical shear in the middle of the nearly dry-adiabatic layer (Figs. 4b and 6g). Although vertical shear increased in the stable layer at $t = 125$ min, vertical wind speed did not increase notably in the layer (i.e., KH instability was not activated) because the magnified stability kept the Richardson number sufficiently larger than 0.25 (data not shown).

If the convective cells appearing after $t = 77$ min are Rayleigh–Bénard convection, they must be activated when the Rayleigh number exceeds the critical value. The Rayleigh number $Ra$ in the atmosphere is obtained from the following equation (Weckwerth et al. 1999):

\[
Ra = \frac{g \beta D T^3}{v^4}
\]
where $g$ is gravitational acceleration; $\bar{\theta}$ and $\Delta \theta$ are the mean and differential potential temperature across an absolutely unstable layer, respectively; $h$ is the depth of the absolutely unstable layer; and $K_M$ and $K_H$ are the eddy diffusivities for momentum and heat, respectively. Here, $K_M = K_H = 30 \text{ m}^2 \text{s}^{-1}$ was used according to Krishnamurti (1975) and Helfand and Kalnay (1983). Since the critical Rayleigh number $R_{ac}$ varies with boundary conditions, a value for both free boundaries, $R_{ac} = 657.5$ (Emanuel 1994), was applied here. The Rayleigh number first has a nonzero value ($Ra = 657.5$) at $t = 55$ min (Fig. 11), which means an absolutely unstable layer appeared below the cloud base ($z = 3900$–$4000$ m) (Fig. 9a); however, no convection had occurred until $t = 77$ min when the Rayleigh number first exceeded the critical value. The Rayleigh number peaked at $t = 93$ min ($Ra \approx 3200$), then rapidly dropped to zero at $t = 104$ min when the vertical wind speed peaked (Fig. 10g). This confirms that the convection cells are activated when $Ra > R_{ac}$, and that the cells manifest Rayleigh–Bénard convection.

The above-mentioned simulation results agree with the calculations of Harris (1977), who showed that an absolutely unstable layer was produced and a stable layer was magnified by cooling from the sublimation of solid precipitation below cloud base. In addition we obtained other results: Rayleigh–Bénard convection appeared in the absolutely unstable layer with a virga-like distribution of snow, then a nearly dry-adiabatic layer with uniform horizontal wind formed below the stable layer because of the convective mixing, and vertical shear increased at the top and the bottom of the nearly dry-adiabatic layer because of the vertical redistribution of momentum. The simulated horizontal and vertical scales of the convection were about 800–1000 and 450 m, respectively, and the vertical wind speeds ranged from about $-3.8$ to $+3.1 \text{ m s}^{-1}$, which were suitable for relatively strong turbulence for large aircraft. These scales were also in good agreement with the observation of holes and stalactites in Harris (1977).
fall velocity. Because the bottoms of mixed-phase alto-cumulus clouds in the midlatitudes consist of ice water content rather than liquid water content (Carey et al. 2008), clouds with base temperatures less than 0°C are adequate for MCT. Since the derived equivalent vertical gust velocities of 4.5–9 m s^{-1} correspond to heavy turbulence (WMO 2003), our sensitivity experiments showed that cloud-base temperatures between approximately 0°C and -15°C with sufficiently dry subcloud conditions would be suitable for generating relatively strong turbulence. These results agree with observations that temperatures near the frontal zone (the cloud base) were from about -3°C to -15°C (Figs. 4a, 5, and 6b), and that the subcloud conditions were dry (Fig. 4a). According to Lewis (1991), the height range of midlevel clouds is approximately 2–7 km, which corresponds to from +2°C to -30.5°C in the International Standard Atmosphere. Therefore, MCT usually will be found beneath midlevel clouds rather than beneath high- or low-level clouds.

When the initial lapse rate beneath the stable layer was set to 6.5 K km^{-1} (Table 2), the differences became smaller than those for an initial lapse rate beneath the stable layer of 8 K km^{-1} (Table 1), indicating that MCT occurs more readily when ambient lapse rates in subcloud layers are close to the dry-adiabatic lapse rate.

**Fig. 12.** As in Fig. 10, but for the results of the simulation with no vertical shear.
4. Discussion

As described in section 1, the generation mechanism and ambient conditions for mammatus clouds look similar to those for MCT since the observed soundings associated with mammatus had a dry-adiabatic subcloud layer with low relative humidity and the largest term producing the negative buoyancy was the cooling due to sublimation (Kanak et al. 2008). However, there are some differences between mammatus clouds and MCT. First, MCT probably associates with virga rather than mammatus, as shown in Fig. 8. Second, the cloud-base temperatures shown in Kanak et al. (2008) are colder than $-20^\circ C$ for all soundings with most about $-20^\circ C$ or colder, which is somewhat cold for MCT occurrence. Third, according to Stith (1995), vertical velocities in mammatus clouds observed by aircraft ranged from about $-2$ to $+1$ m s$^{-1}$, the order of which agrees with many other observations (Table 2 in Kanak et al. 2008), which correspond to light turbulence (WMO 2003) and are inconsistent with frequent moderate turbulence occurrence in MCT. Fourth, mammatus typically can be seen beneath cumulonimbus cirrus anvils with time periods of 15 min to as much as a few hours (Schultz et al. 2006), while MCT typically will be found beneath midlevel clouds and sometimes continues for more than several hours (Fig. 5).

The environmental conditions for generating dry microbursts appear to be similar to those for generating MCT except for the temperature profile below cloud bases. Environments favorable for dry microbursts have a deep, dry-adiabatic, and very dry subcloud layer from the surface to a midlevel cloud (Wakimoto 1985), while for MCT, dry-adiabatic layers can usually be observed between cloud base and a few thousand meters below cloud base but they do not reach the ground (Figs. 4a and 6). The other characteristics of dry microbursts are quite similar to those of MCT. For example, the primary forcing mechanism is the sublimation of snowflake aggregates, dry microbursts are accompanied by little or no rain at the ground, and the downdrafts are usually present with virga, which consists of ice particles rather than raindrops. Though vertical wind velocities differ significantly between dry microbursts and MCT, MCT is very similar to dry microbursts that do not reach the ground.

5. Summary

We presented a typical case of midlevel cloud-base turbulence by using an upper-air sounding and aircraft
observations. To clarify the mechanism of its generation, we performed high-resolution three-dimensional numerical simulations with idealized initial conditions. These simulations showed that MCT was caused by Rayleigh–Bénard convection resulting from cooling due to sublimation of snow. The simulated cloud base was accompanied by a virga-like distribution of snow (Fig. 13 gives an example of virga). The convection formed a nearly dry-adiabatic layer with a homogeneous wind profile below the cloud base, which caused the vertical shear to increase both at the top and the bottom of this nearly dry-adiabatic layer and to decrease in the middle of this nearly dry-adiabatic layer. These results were consistent with the upper-air sounding and aircraft observations. Our sensitivity experiments showed appropriate conditions to bring about MCT: 1) cloud-base temperature is between about 0° and −15°C, 2) relative humidity in subcloud layers is sufficiently low, and 3) stability in subcloud layers is weak (the lapse rate is already close to the dry-adiabatic lapse rate). Vertical shear was not necessary for the generation of MCT. Suitable cloud-base temperatures for the generation of MCT are found in the midlevels of the atmosphere.

The simulated Rayleigh–Bénard convection had the property that the horizontal scale was about 800–1000 m and the variations of vertical wind velocity were up to about 7 m s\(^{-1}\). These scales are suitable for relatively strong turbulence for large aircraft. The generation mechanism of turbulence presented in this study is an important cause of in-flight turbulence that affects aircraft.

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