Some Influences of Background Flow Conditions on the Generation of Turbulence due to Gravity Wave Breaking above Deep Convection

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(Manuscript received 22 May 2007, in final form 10 March 2008)

ABSTRACT

Deep moist convection generates turbulence in the clear air above and around developing clouds, penetrating convective updrafts and mature thunderstorms. This turbulence can be due to shearing instabilities caused by strong flow deformations near the cloud top, and also to breaking gravity waves generated by cloud–environment interactions. Turbulence above and around deep convection is an important safety issue for aviation, and improved understanding of the conditions that lead to out-of-cloud turbulence formation may result in better turbulence avoidance guidelines or forecasting capabilities. In this study, a series of high-resolution two- and three-dimensional model simulations of a severe thunderstorm are conducted to examine the sensitivity of above-cloud turbulence to a variety of background flow conditions—in particular, the above-cloud wind shear and static stability. Shortly after the initial convective overshoot, the above-cloud turbulence and mixing are caused by local instabilities in the vicinity of the cloud interfacial boundary. At later times, when the convection is more mature, gravity wave breaking farther aloft dominates the turbulence generation. This wave breaking is caused by critical-level interactions, where the height of the critical level is controlled by the above-cloud wind shear. The strength of the above-cloud wind shear has a strong influence on the occurrence and intensity of above-cloud turbulence, with intermediate shears generating more extensive regions of turbulence, and strong shear conditions producing the most intense turbulence. Also, more stable above-cloud environments are less prone to turbulence than less stable situations. Among other things, these results highlight deficiencies in current turbulence avoidance guidelines in use by the aviation industry.

1. Introduction

Deep convective clouds interact with their environment by inducing organized circulations of inflow, outflow, subsidence, and ascent, and by generating internal gravity waves that can propagate large distances horizontally and vertically from their originating source. Instabilities near the cloud top (CT) (e.g., Grabowski and Clark 1991) and the breakdown of convectively induced gravity waves (e.g., Haman 1962; Wang 2003; Lane and Sharman 2006) can induce turbulence and mixing in the upper troposphere and lower stratosphere. This turbulence can be responsible for creating well-mixed layers that distribute chemical constituents vertically (e.g., Moustaoui et al. 2004), can initiate cirrus clouds (Wang 2004), and poses a safety issue for aviation (e.g., Pantley and Lester 1990; Prophet 1970). In fact, a study of severe turbulence incidents in the United States by Kaplan et al. (2005) showed that 86% of the turbulence encounters examined occurred within 100 km of deep convection, and turbulence attributed directly to clouds or convection constituted the second-leading cause of turbulence.

Although progress has been made in the prediction of aircraft-scale turbulence associated with jet streams

* The National Center for Atmospheric Research is sponsored by the National Science Foundation.

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DOI: 10.1175/2008JAMC1787.1

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and upper-level frontal systems (see Sharman et al. 2006 and references therein), and explicit predictions of mountain-wave-induced turbulence show promise (e.g., Sharman et al. 2004; Doyle et al. 2005), the prediction of turbulence associated with convective clouds is much more difficult, in part because the scales of the forcing are much smaller and more transient, and in part because the nature of the turbulence generation processes remains poorly understood. Nevertheless, it is recognized that there are two distinct turbulence generation mechanisms associated with cloud: (i) in-cloud convective instabilities, and (ii) out-of-cloud turbulence generated by the cloud’s presence (e.g., Lester 1993).

The prediction of in-cloud turbulence requires the prediction of the occurrence and intensity of deep convection. Given the highly spatially and temporally varying character of instabilities within convective clouds, skillful numerical prediction of in-cloud turbulence is unlikely in the foreseeable future, although aircraft may easily avoid in-cloud turbulence by simply diverting away from regions of visible or radar-inferred cloud boundaries. The prediction of out-of-cloud turbulence, however, requires a prediction of convective development, and a determination as to whether these convective clouds will generate turbulence (either due to local instabilities near the cloud boundary or due to the breakdown of gravity waves generated by the cloud itself). However, the short time scale of predictability of deep convection undermines these tasks, making deterministic forecasts of these events extremely difficult. This predictability limit suggests that the most feasible approach to predicting and possibly avoiding out-of-cloud turbulence is to examine (analyzed or forecast) background flow conditions to determine the likelihood, extent, and intensity of turbulence, in the event that a convective system were to develop.

Current Federal Aviation Administration (FAA) guidelines suggest avoiding severe thunderstorms by 20 miles (32 km) in the horizontal and flying above thunderstorms by at least 1000 ft (305 m) for every 10 kt (5 m s\(^{-1}\)) of cloud-top wind speed (FAA 2008). This recommended vertical separation does not take into account the vertical wind shear in the vicinity of the cloud top nor the background static stability, both of which should play an important role in the breakdown of gravity waves and other instabilities (Lane et al. 2003). With this in mind, the aim of this study is to examine the sensitivity of breaking gravity waves directly above deep convection to changes in these background flow conditions to better understand the important environmental factors contributing to the generation of out-of-cloud turbulence.

The focus of this study is turbulence in the clear air above cloud, a known aviation hazard (Pantley and Lester 1990; Prophet 1970). The generation of gravity waves by convection and their propagation into the stratosphere is well known (e.g., Fovell et al. 1992; Lane et al. 2001), and in a numerical modeling study of a severe turbulence encounter that occurred directly above deep convection, the breakdown of these convectively induced gravity waves was found to be the most likely explanation for the turbulence (Lane et al. 2003). In the Lane et al. case study, the convectively induced gravity waves had short periods (10–15 min), short horizontal wavelengths (5–10 km), and therefore relatively slow horizontal phase speeds (5–15 m s\(^{-1}\)). Lane et al. showed that when these relatively short-scale, high-frequency waves propagate vertically in an environment with moderate above-cloud wind shear, the waves can either decrease in amplitude with height (i.e., they are evanescent) or as is well-known (e.g., Booker and Bretherton 1967; Fritts 1982), may interact with a critical level [the level \(z_c\), where \(U(z_c) - c = 0\)] and break. Here \(U(z)\) is the background wind speed, a function of height \(z\), in an arbitrary reference frame, and \(c = \omega/k\) is the wave phase speed in that reference frame; \(\omega\) is the wave frequency and \(k\) is the horizontal wavenumber.

As an example, guided by the Lane et al. (2003) study, consider waves that are generated with a horizontal wavelength of 5 km (\(|k| = 1.26 \times 10^{-3}\) rad m\(^{-1}\)) and a frequency \(\omega\) of 0.007 rad s\(^{-1}\) (in a frame of reference moving with the top of a convective cloud). The corresponding phase speed is \(c = 5.6\) m s\(^{-1}\). Assume waves propagate in both directions with this speed, and the background wind speed at the source is zero. If there is a change in background wind speed \(U\) above the cloud top of 5.6 m s\(^{-1}\), \(U - c\) will equal zero for those waves propagating in the same direction as that change (i.e., downshear). This results in a critical level for the downshear gravity waves, and possible wave breaking. Also, if we use a typical value of the stratospheric Brunt–Väisälä frequency of \(N = 0.02\) rad s\(^{-1}\), \(N/k = 15.9\) m s\(^{-1}\). Therefore, \(|U - c| \approx N/k\) for those waves propagating in the opposite direction to the change in wind (i.e., upshear) if the change in wind speed is 10.3 m s\(^{-1}\). At the height of this change in wind the waves would become evanescent (as defined by the nonhydrostatic dispersion relation). These processes are summarized schematically in Fig. 1, and the reader is referred to Lane et al. (2003) and Lane and Sharman (2006) for more details.

It is the breakdown of the downshear-propagating gravity waves that probably makes the most important contribution to turbulence generation above convect-
tion. In this study we extend the Lane et al. (2003) results to examine the effect of changing the strength of the above-cloud wind shear and static stability on the breakdown of gravity waves in the lower stratosphere, and the resultant influence of these changes on the extent and intensity of turbulence and mixing produced by the instabilities. This is the first such study to examine these sensitivities systematically, allowing an assessment of existing FAA guidelines and the important underlying processes. Understanding these processes is of crucial importance for future turbulence avoidance strategies and other aviation safety applications. Moreover, understanding the breakdown of gravity waves above convection has fundamental importance also, because the dissipation of these waves plays an important role in the momentum budget of the middle atmosphere (see Fritts and Alexander 2003).

This paper is arranged as follows: section 2 outlines the modeling approach. Section 3 describes a series of two-dimensional (2D) simulations of deep convection and quantifies the instability induced by cloud-top instabilities and gravity wave breakdowns. These simulations investigate the influence of above-cloud wind shear and static stability on the occurrence and intensity of the above-cloud turbulence and mixing. In section 4, some three-dimensional (3D) simulations are conducted to determine how well the 2D simulations represent a fully 3D flow. Section 5 discusses and interprets the results of the simulations, which are summarized in section 6.

2. Modeling approach

The goal of this study is to explore the sensitivity of the above-cloud turbulence to background flow conditions that may play a role in the turbulence generation process. This sensitivity may include such factors as the intensity of the thunderstorm [as measured, e.g., by the convective available potential energy (CAPE)] and environmental conditions that may influence the generation, propagation, and breakdown of the convectively induced gravity waves above the thunderstorm [e.g., wind speed, wind shear (speed and directional), and static stability above the cloud top]. However, given the large parameter space that must be covered to obtain a complete understanding of the important turbulence generation processes, a tremendous number of simulations would have to be performed. Therefore this study must necessarily be limited in its scope, and concentration will be on those environmental conditions that would most likely affect the propagation and breakdown of convectively induced gravity waves in the

![Figure 1. Schematic of gravity wave propagation and breakdown above deep convection in an environment with negative above-cloud wind shear, in a reference frame moving with the cloud-top wind. If the change in wind speed is sufficiently large the waves with negative phase speed will encounter a critical layer and break down into turbulence, while the waves with positive phase speeds will become evanescent. Typical values of phase speeds and horizontal wavelengths are shown.](image-url)
stability (as measured by the Brunt–Väisälä frequency, $N$), where $z$ is height). To further simplify the analyses, we restrict the numerical studies to relatively simple analytical-form profiles of stratospheric background wind and stability.

Even with these limitations, conducting all of the necessary simulations at the required model resolution in three dimensions is prohibitively computationally expensive. Therefore, most of the simulations presented use a two-dimensional, slab-symmetric model configuration. To determine if these 2D simulations produce results that are representative of a 3D flow, a few 3D model simulations will also be presented for comparison purposes in section 4.

Model description and experimental design

1) Numerical model

The numerical model used in this study is the cloud-scale model originally developed by Clark (1977), documented in Clark et al. (1996), and used by Lane et al. (2003) in the convectively induced gravity waves and turbulence case study mentioned above. The model is a second-order finite-difference approximation to the nonhydrostatic, anelastic equations of motion. Water vapor, condensed water (cloud and rain), and ice variables are treated explicitly in the model, utilizing a combination of the Kessler (1969) and the Koenig and Murray (1976) microphysics parameterizations.

Of particular importance to this study is the treatment of subgrid-scale turbulence processes. Here subgrid-scale mixing is parameterized using a first-order Smagorinsky closure (Smagorinsky 1963; Lilly 1962). In this scheme, the subgrid-scale mixing coefficient of momentum, $K_M$, is zero everywhere unless the local Richardson number $Ri$ is less than unity. In the latter case, $K_M$ is

$$K_M = (C\Delta)^2|\text{Def}| \sqrt{1-Ri}, \quad (1)$$

where, consistent with an assumed Kolmogorov turbulence spectrum, $C = 0.2$, $\Delta$ is the geometric average of the grid increments, and $\text{Def}$ is the total deformation defined using

$$\text{Def}^2 = \frac{1}{2} \sum_i \sum_j D_{i,j}^2 \quad (2a)$$

where

$$D_{i,j} = \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} - 2\delta_{i,j} \frac{\partial u_k}{\partial x_k} \quad (2b)$$

is the deformation tensor (written in standard tensor notation). The Richardson number is

$$Ri = \frac{N^2}{\text{Def}^2}, \quad (3)$$

where $N$ is the Brunt–Väisälä frequency. The eddy Prandtl number is assumed to be unity; thus $K_H = K_M$, where $K_H$ is the eddy diffusion coefficient for heat. Using the usual turbulence closure assumptions, $K_M$ can be related to a subgrid-scale turbulent kinetic energy (TKE) using (e.g., Deardorff 1980)

$$\text{TKE}^{1/2} \approx 10K_M/\Delta. \quad (4)$$

For all simulations the model is configured to include a Rayleigh friction absorbing layer near the model top to mitigate the reflection of vertically propagating disturbances off the model lid. Cyclic lateral boundary conditions are used, which allows consistent comparisons to be made between cases; the performance of open lateral boundaries varies with the wind profile. Further, most analyses are performed at early times, and therefore transient disturbances do not have time to reach the boundaries and propagate back into the interior of the domain.

2) Background profiles

The 0000 UTC 11 July 1997 Bismarck, North Dakota, sounding (BIS) is used to provide background thermodynamic fields for all model simulations and is shown in Fig. 2. (Above the top of the observed sounding, 33 km, the background temperature and relative humidity in the model are constant with height and equal to the uppermost observed values.) This sounding was also used in the 2D simulations described in Lane et al. (2003) and in the 3D simulation described in Lane and Sharman (2006), which both used a unidirectional wind defined by the 215° component of the observed wind. In this study, the background wind profile below $z = 5$-km height is the same as the 215° component from the BIS sounding. Above 5 km, a jetlike wind profile is used:

$$U(z) = U_{\text{max}}e^{-(z-z_m)/\sigma^2}, \quad (5)$$

where $z_m$ is the level of the maximum wind $U_{\text{max}}$. Vertical interpolation is used surrounding $z = 5$ km to ensure a smooth transition between the observed and prescribed wind. The parameters $U_{\text{max}}$ and $\sigma$ are used to vary the maximum background wind speed and the width of the jet (see Fig. 3). Therefore, positive shear exists below $z = z_m$ with negative shears above this level. In all cases to be presented, $z_m = 12$ km (i.e., the level of the jet maximum is just above the approximate
tropopause height and most of the lower stratosphere is in the negative shear region. Parameter settings of $U_{\text{max}}/H_{1100} = 15 \text{ m s}^{-1}$ and $\sigma = 3 \text{ km}$ produce a wind profile very similar to that observed in the BIS sounding. Most simulations use the thermodynamic profile from the BIS sounding, and wind profile parameter settings for these simulations are shown in Table 1. Also, the effect of lower-stratospheric stability is investigated in a few simulations by setting the Brunt–Väisälä frequency above 11.5 km, $N_s$, to a constant value ranging from 0.015 to 0.025 $\text{s}^{-1}$. In the cases that use the BIS sounding, $N_s$ is approximately 0.022 $\text{s}^{-1}$ but is not exactly constant with height.

3) **Turbulence estimation**

To quantify the turbulence generated above the convection, the following method is used. First, the cloud boundary is identified by the 0.1 g kg$^{-1}$ contour of total cloud loading (the sum of the mixing ratios of cloud condensate and the two ice types). Model columns with a cloud-top altitude greater than 7 km, with at least 1 km of contiguous cloud below that top, are identified; these identified columns satisfy our criterion that the cloud is sufficiently deep and thick. (This somewhat arbitrary choice of 7 km was made because it effectively separates those deep clouds that eventually reach the upper troposphere from those shallow clouds that reside in the boundary layer and lower troposphere.) Only the grid cells above these identified columns are used to quantify the above-cloud turbulence; namely, those grid cells that are above cloud and contain $K_M > 1 \text{ m}^2 \text{s}^{-1}$ are used to identify regions of wave breaking or instability. The 1 m$^2$ s$^{-1}$ threshold is admittedly arbitrary, but the results to be presented are fairly insensitive to the choice of this or smaller thresholds applied to $K_M$. In the model’s subgrid closure, $K_M$ is only nonzero when the Richardson number is less than unity (see, e.g., Miles 1986), and therefore this criterion identifies regions that may be subject to wave-induced overturning or shearing instabilities (i.e., dynamic or static instabilities). For each model simulation, the number of grid cells that are above cloud and contain nonzero $K_M$
are separated into two groups. The first group (CT–CT500) contains those grid cells within 500 m of the cloud top, while the second group (CT500–25km) contains those grid cells between 500 m above cloud top and 25-km altitude. The number of grid cells containing nonzero $K_M$ in each group is multiplied by $\Delta x \Delta z$ to obtain the area of instability.

The total area of above-cloud instability is represented by the sum of the areas from CT–CT500 and CT500–25km. However, the CT–CT500 group may contain a combination of instabilities due to wave breaking and instabilities due to the presence of the cloud boundary, such as the cloud-interfacial instability (Grabowski and Clark 1991). The upper group (CT500–25km) contains parts of the flow that are sufficiently far from the uppermost cloud boundary so that the regions of instability can be assumed to be mostly due to convectively induced gravity wave breaking.

The above method will not identify areas of turbulence laterally adjacent to (i.e., not directly above) the convection. However, in almost every case examined, the majority of nonzero $K_M$ regions occur above the cloud and there are no substantial areas missed by this method. The exceptions to this statement are a few cases with the highest wind shear that feature contiguous regions of mixing adjacent to the cloud. These regions are due to organized cloud circulations interacting with the layers of low background Richardson number.

3. Two-dimensional numerical modeling results

Because of the relative ease and low cost of performing 2D simulations, a substantial number of very high-resolution runs can be performed with varying $U_{\text{max}}$, $\sigma$, and $N_s$. In all, a total of 24 simulations are performed (Table 1). Each simulation uses the same model domain configuration. The domain is 250 km wide and 40 km deep, with 2000 grid points in the horizontal and 400 grid points in the vertical directions. Therefore, the horizontal grid spacing is 125 m and the vertical grid spacing is 100 m. The time step is 1 s. This resolution is probably sufficient to resolve the spectrum of gravity waves generated by the storm (see Lane and Knievel 2005). The Rayleigh absorber depth is 15 km. The simulations are initialized with a rectangular bubble in the center of the model domain at the surface. The bubble is 10 km wide and 2 km high and takes the form of a 2-K perturbation in the potential temperature. Such an initialization allows the convection to develop similarly in each case, which aids in their intercomparisons.

a. The baseline case

The baseline simulation uses values of $U_{\text{max}} = 15$ m s$^{-1}$, $\sigma = 3$ km, and the thermodynamic profile observed by the BIS sounding. The simulation produces a modeled cloud that is qualitatively similar to that produced in Lane et al. (2003), with slight differences in the

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10 km wide and 2 km high and takes the form of a 2-K perturbation in the potential temperature. Such an initialization allows the convection to develop similarly in each case, which aids in their intercomparisons.
details of the wave breaking and cloud width because of the idealized wind profile and the different model resolutions. Figure 4 shows the cloud, potential temperature, and eddy-mixing coefficient at two times during the evolution of the modeled convection for the baseline case. At 60 min (Fig. 4a), the modeled convection has reached the tropopause (~11.5 km) and has broadened downstream to form a well-defined anvil. Above the deep convection, in the lower stratosphere, gravity waves are evident in the potential temperature field. These waves have horizontal wavelengths of 5–15 km. Similar to previous studies (e.g., Lane et al. 2003; Lane and Sharman 2006), the gravity waves propagating in the same direction as the above-cloud shear vector (i.e., in the negative x direction) break in the lowest 4–5 km of the stratosphere. This breakdown shows turbulence and mixing from the cloud top to about 16-km altitude. At 120 min (Fig. 4b), the cloud has broadened to be
greater than 100 km wide, and the wave breaking has continued, increasing in horizontal extent and depth. However, the extent of turbulence and mixing near the cloud top has decreased, and the region of wave breaking has become disconnected from the cloud top. These results are in agreement with the simulations described in Lane et al. (2003), which analyzed in detail the effect of gravity wave–critical level interactions on the gravity wave breakdown and production of turbulence. Consistent with the earlier results, this simulation shows that the areas of wave breakdown are confined to the downshear regions above the cloud as the waves generated by the convection approach their critical levels. Since there is a spectrum of waves generated, each wave component of the spectrum will have a different critical level. However, because of this and nonlinear effects, the breakdown and turbulence production region is smeared out over a layer, and does not occur preferentially at one single level [see Lane et al. (2003) and other references to gravity wave–critical level interactions (e.g., Booker and Bretherton 1967; Fritts 1982)].

Figure 5 quantifies the evolution of above-cloud instabilities in terms of the area of nonzero above-cloud $K_{N}$ to take into account the increase in the cloud width with time, Fig. 5b shows the area of instability normalized by the cloud width. Figure 5b shows that at early times, the area of instability near the cloud top (CT–CT500) is maximized, and shows a decreasing trend with time. On the other hand, the normalized area of instability from CT500–25km shows an increasing trend with time, with a rapid increase in normalized area between 30 and 60 min. Between 60 and 120 min, the normalized area continues to increase but at a slower rate than at earlier times, and after 120 min the changes are small.

To illustrate the influence of the cloud width normalization, the evolution of the (raw) areas of instability for the baseline case is also shown (Fig. 5a). First consider the area closest to the cloud top (CT–CT500); the total area of instability near the cloud top is fairly constant with time (i.e., it does not show the decreasing trend that the normalized area shows). Therefore, as time continues, the total area of instability near the cloud top is approximately constant, but the area per unit length of cloud decreases with time. This is an intuitive result, because at earlier times when the cloud is most intense, the convective overshoot is maximized, as are flow deformations that are likely to cause turbulence and mixing near the cloud top. In the upper layers (CT500–25km), however, the total area of instability increases almost linearly with time, and shows the same general trend as the normalized cloud area, except the normalized cloud area shows a weaker trend due to the increase in cloud width with time. If the instability in this upper region (CT500–25km) is assumed to be due
to gravity wave breaking, the baseline simulation shows that the gravity wave breaking develops about an hour into the model simulation with the total area of breaking increasing with time.

The use of the normalized areas of instability is necessary because of the geometric method being used to quantify the turbulence, and is also particularly relevant for aviation applications; the larger the normalized area of instability the more likely it is that turbulence will be encountered above cloud. Furthermore, by dividing the normalized areas of instability by the depth of the layer in question (e.g., 0.5 km), the proportion of turbulent air can be estimated. For example, from Fig. 5b it can be inferred that at 30 min, approximately 25% of the volume of air in the lowest 500 m above cloud top is turbulent, whereas at 120 min only about 5% is turbulent in this layer.

b. The influence of jet width

To investigate the influence of the width of the jet in the background wind profile, a series of nine 2D simulations are completed with different values of the jet width parameter, \( \sigma \), with a fixed maximum wind speed of the jet \( U_{\text{max}} = 15 \text{ m s}^{-1} \) (these simulations are listed as cases 3 and 5–12 in Table 1). Small values of \( \sigma \) represent large wind shears, and large values of \( \sigma \) represent weak wind shears. Using the arguments given in section 1 above, large values of \( \sigma \) increase the height of the critical level, which is the primary cause of the wave breaking in the baseline case. Similarly, small values of \( \sigma \) cause the critical level to be closer to the cloud top.

Simulated fields at 60 min from two cases that represent extreme values of \( \sigma \) are shown in Fig. 6, which can be compared to an intermediate value of \( \sigma \) (Fig. 4a, the baseline case). The case with the weakest wind shear (Fig. 6a, \( \sigma = 6 \text{ km} \)) shows a wave field that is close to symmetric, with less evidence of breaking than the baseline case and relatively large amplitude waves aloft. The case with the narrowest jet and hence strongest wind shear (Fig. 6b, \( \sigma = 0.75 \text{ km} \)) shows a small area of wave breaking that is more localized and at lower altitudes than in the baseline case. This case also shows evidence of large amplitude propagating waves above the jet. This strong wind shear case also includes a contiguous region of nonzero \( K_M \) close to the cloud top that is collocated with the region of strongest wind shear in the jet. This region of mixing is due to the enhancement of the background wind shear by the cloud circulation to the point that the instability threshold (\( \text{Ri} < 1 \)) is reached. This region is located within the CT–CT500 group and does not contribute to the quantification of the wave breaking farther aloft.

The results from the nine simulations with different jet width parameters and \( U_{\text{max}} = 15 \text{ m s}^{-1} \) are summarized in Figs. 7a and 7b, which show the normalized area of instability at 60 and 120 min. Figures 7a,b show that the area of instability near the cloud top (CT–CT500) exhibits an inverse relation to the jet width parameter, \( \sigma \). Equivalently, the area of instability near the cloud boundary is largest for the strongest wind shears, and smallest for the weakest wind shears. As the wind shear increases, the background Richardson number is reduced, which allows flow deformation near the cloud boundary to generate mixing more readily. However, the normalized area of mixing at upper levels (CT500–25km) does not show the same relationship as the near-cloud mixing; rather the area of breaking is maximized at intermediate values of \( \sigma \) (intermediate shear). Both strong and weak shear cases show similar values of normalized area, which is approximately half of the maximum value that occurs with intermediate shear. A comparison of Figs. 7a and 7b shows that these relationships between normalized area and vertical shear are insensitive to time.

The above analysis is repeated for the seven cases from Table 1 that use \( U_{\text{max}} = 22.5 \text{ m s}^{-1} \), and is shown in Figs. 7c and 7d. Similar to the 15 m s\(^{-1}\) cases the normalized area of instability at 60 min shows a maximum area of instability at intermediate values of \( \sigma \). However, at 120 min (Fig. 7d) the normalized area of instability does not vary substantially with \( \sigma \) (except at \( \sigma = 6 \text{ km} \)). Unfortunately, at these later times and intermediate \( \sigma \) values, the \( U_{\text{max}} = 22.5 \text{ m s}^{-1} \) simulations feature some additional cloud development (not shown). This development does not appear to contribute to the turbulence area, but does increase the cloud width that is used for the normalization. This additional cloud width effectively penalizes the raw turbulence areas at these intermediate shears, removing any maxima in turbulence area that should be expected.

As mentioned above, when the wind profile is modified it causes a change in the evolution of the cloud, the width of the cloud, and the characteristics of the source of gravity waves. It is impossible to isolate the effects of these changes on the details of the wave breaking, and therefore some influence of this effect remains in the results. However, by comparing the normalized areas of turbulence intensity, some of these influences are removed, making the comparisons more robust. In any event, the bottom line from the aviation safety perspective is unaffected; to wit, intermediate shear values are more conducive to turbulence generation above cloud than relatively large or small shear values. Operationally, incorporating the cloud width into algorithms that assess above-cloud turbulence potential would be easy.
to implement since that measure is readily available from satellite imagery; however, other specific influences of wind shear on the cloud dynamics would be much more difficult to ascertain operationally.

c. The influence of wind shear

The previous comparisons of jet width examined cases with a fixed maximum wind speed of the jet \(U_{\text{max}} = 15 \text{ m s}^{-1}, 22.5 \text{ m s}^{-1}\), and essentially examined the sensitivity of the results to wind shear. These comparisons are important for aviation applications because current turbulence avoidance guidelines are stated in terms of cloud-top wind speed only, while the results presented in the previous subsection illustrate that for the same cloud-top wind speed there can be substantial variation in turbulence occurrence. Theoretical arguments (e.g., Lane et al. 2003) show that wind...
shear, or more correctly a change in wind speed relative to the cloud top, is necessary to induce wave breaking via critical-level interactions.

All simulations use the wind profile defined by Eq. (5), for which the wind shear is

\[ \frac{dU}{dz} = \frac{-2(z - z_m)}{\sigma^2} U_{max} e^{-\frac{(z - z_m)^2}{\sigma^2}}. \]  

Note that this has a maximum magnitude at \( z = z_m \pm \sigma \sqrt{2}/2 \) km of

![Fig. 7. Normalized area of instability from CT–CT500 (squares) and CT500–25km (crossed squares), at (a), (c) 60 and (b), (d) 120 min. Panels (a) and (b) are for cases with \( U_{max} = 15 \text{ m s}^{-1} \); (c) and (d) are for cases with \( U_{max} = 22.5 \text{ m s}^{-1} \). See Table 1 for details.](image)
Therefore, the maximum wind shear in the sounding is proportional to \( \frac{U_{\text{max}}}{\sigma} \), and for the remainder of this paper \( U_{\text{max}}/\sigma \) is referred to as the “wind shear parameter.” In this section, the sensitivity of the occurrence of instability and turbulence to this wind shear parameter is evaluated using simulations that vary both the maximum wind speed \( U_{\text{max}} \) and the jet width parameter \( \sigma \). A further 10 simulations (total of 19) are conducted as summarized in Table 1.

First, consider cases 1–4 (Table 1), which vary \( U_{\text{max}} \) from 0 to 22.5 \( \text{m s}^{-1} \) while holding \( \sigma \) constant (3 km). At 120 min these four cases feature normalized areas of instability from CT500–25km of 0.022, 0.081, 0.287, and 0.293 \((\text{km}^2)/\text{km}\), respectively (not shown) that progressively increase with \( U_{\text{max}} \). Therefore, the normalized area at intermediate values of \( \sigma \) and \( U_{\text{max}} = 15 \text{ m s}^{-1} \) (Fig. 7b) is also larger than weaker shear cases with slower maximum wind speeds. In other words, large normalized areas of instability are favored by larger maximum wind speeds but also by intermediate values of the wind shear. However, this is not necessarily an absolute maximum in \( (U_{\text{max}}, \sigma) \) space (e.g., Fig. 7d). Increasing \( U_{\text{max}} \) further would allow investigation of this peak in more detail; however, in such cases, the maximum wind speed is too large and the cloud tops become sheared off. This results in large differences in the cloud properties compared to weaker shear cases, making further comparisons nonsensical.

The normalized areas of instability at 120 min (the time after which the normalized area of instability is fairly constant; cf. Fig. 5) from the 19 simulations for CT–CT500 and CT500–25km are shown in Fig. 8 versus the wind shear parameter \( U_{\text{max}}/\sigma \). Figure 8a illustrates that the normalized area of instability near the cloud top (CT–CT500) shows a strong relationship with the wind shear, with larger normalized areas for larger wind shears. As mentioned earlier, this result is not surprising. While there is a clear positive trend, at low wind shears there is a noticeable variance in the normalized area values, possibly due to the clouds being at slightly different stages of their evolution (cf. Fig. 4).

Figure 8b shows the normalized instability area versus wind shear for the region CT500–25km, which represents the breakdown of gravity waves. Normalized area increases with increasing wind shear until approximately \( U_{\text{max}}/\sigma = 5 \text{ m s}^{-1} \text{ km}^{-1} \). At shear values larger than 5 \text{ m s}^{-1} \text{ km}^{-1} there is substantial spread in the normalized area values, and there is a suggestion of a decreasing trend at these larger shears and therefore an optimum value of the shear parameter at approximately 5 \text{ m s}^{-1} \text{ km}^{-1}. This result can be understood by considering potential gravity wave–critical level interactions that will be discussed in section 5. The variance of the

\[
\frac{dU}{dz} \bigg|_{z=z_m-\sigma\sqrt{2}} = \pm \sqrt{2} \frac{U_{\text{max}}}{\sigma} e^{-1/2}.
\]
results at larger shears is caused by the variance in cloud evolution and wave generation that are a result of the different environmental wind profiles. Furthermore, variations in \( U_{\text{max}} \) cause larger changes in the properties of the cloud in the upper troposphere in comparison to changes in \( \sigma \) only—these sensitivities add to the variance between all of the 19 cases.

While the area of instability provides guidance as to the occurrence and extent of turbulence, which includes mixing due to breaking gravity waves aloft and instabilities near the cloud boundary, it does not provide an estimate of the intensity of the breaking. To estimate the intensity of the subgrid-scale mixing induced by the resolved scale flow, the eddy mixing coefficient \( K_M \) is converted to a subgrid TKE using Eq. (4). The subgrid TKE is then integrated from the cloud top to 25 km to obtain an estimate of turbulence intensity. This integral quantity depends heavily on the assumptions used to convert from \( K_M \) to TKE, and the final integrated value is simply proportional to the integral of \( K_M^2 \). Yet, TKE is used here because it represents a more physically intuitive quantification of turbulence intensity than \( K_M^2 \). The integrated TKE can produce large values for large areas of weak turbulence or for small areas of very strong turbulence. The integrated TKE is shown in Fig. 9a and a normalized value (normalized by the cloud width) is shown in Fig. 9b, both as a function of the wind shear parameter.

Both Figs. 9a and 9b show that the integrated above-cloud TKE generally increases with the strength of the wind shear. This statement is true for all 19 cases, as well as the 9 cases considered in the previous section with \( U_{\text{max}} = 15 \text{ m s}^{-1} \), (shown as triangles in Fig. 9). While there is some spread of results at values of the shear parameter around 10–20 \( \text{ m s}^{-1} \text{ km}^{-1} \), there is no indication of optimum shear conditions at intermediate values, which was seen in the analysis of the area of instability. This analysis presented in Fig. 9 does include instabilities produced near the cloud, but because the integrated TKE is presented at a late time (120 min), turbulence values near the cloud are minimized. In general, the turbulence produced by the above-cloud instabilities increases in intensity as the above-cloud wind shear increases.

d. The influence of lower-stratospheric static stability

To determine the role of lower-stratospheric static stability in controlling gravity wave breaking, a series of simulations are completed with the same settings as the baseline simulation \( (U_{\text{max}} = 15 \text{ m s}^{-1}, \sigma = 3 \text{ km}) \), except using different values of stratospheric static stabil-
ity. Five simulations are conducted with constant stratospheric \((z/\text{H}_{11022} = 11.5 \text{ km})\) stabilities corresponding to Brunt–Väisälä frequencies of 0.015, 0.0175, 0.02, 0.0225, and 0.025 s\(^{-1}\). The normalized areas of instability from these five simulations after 120 min of model time are shown in Fig. 10. The normalized areas of instability from both groups (CT–CT500 and CT500–25km) show a decrease in area as the static stability increases. All else being equal, the background Richardson number will be larger when the static stability is larger, and therefore the flow will be more dynamically stable. Also, a large amplitude wave is more likely to become saturated in conditions that are less stable. The reduction in instability and breaking as the static stability increases is therefore not an unexpected result.

4. Three-dimensional numerical modeling results

The results presented in section 3 all utilize a high-resolution numerical model that is 2D (slab symmetric) and represent a simplification of the real 3D atmosphere, as it does not include 3D cloud geometries and their resultant influence on wave generation, propagation, and breakdown. Although the wave breaking criterion associated with critical-layer interactions for unidirectional flow is unaffected by the extra dimension (e.g., Berkshire 1975), a 2D model does not include 3D effects of cloud–environment interactions, or, importantly for this study, 3D wave dispersion. Therefore, as waves propagate from their source the gravity waves in a 2D model should have larger amplitudes than their 3D counterparts; this larger amplitude should allow waves to break more readily. Also, relative to the generated turbulence fields, the 2D framework does not produce a realistic representation of the turbulent cascade after the waves have begun to break down (e.g., Dörnbrack 1998), nor does it properly represent the mixing at the cloud boundary (e.g., Grosvenor et al. 2007). However, the focus of the present study is on the initiation of the turbulent cascade through wave instability rather than the evolution of the turbulence. With these points in mind, it is important to assess how well the 2D results presented in section 3 represent the processes occurring in 3D. In this section a comparison is made between some two- and three-dimensional simulations to determine whether they are quantitatively similar and whether they show similar relationships to variations in wind shear.

The 3D model is configured with a horizontal grid spacing of 250 m and a vertical grid spacing of 200 m. The domain is 100 km long in the \(x\) direction, 50 km wide in the \(y\) direction, and 36 km deep with a 10-km-deep Rayleigh friction layer at upper levels. To reduce computational costs, simplified microphysics are used employing the Kessler (1969) warm rain parameterization only. The background profile of wind and thermodynamic variables is defined in the same way as the 2D model in section 2, with the (unidirectional wind) directed along the \(x\) axis. Convection is initiated with a warm perturbation that is circular in the horizontal, with a diameter of 10 km and a depth of 2 km (same as the 2D model). This model configuration is the same as that described by Lane and Sharman (2006), except here the model resolution is coarser and an idealized wind profile is used. Available computational resources limit the size and resolution of the 3D model configuration, which makes a comparison with the results presented in section 2 difficult because there are numerous model differences (viz. the model resolution, domain size, and microphysics complexity). To make the comparison between the two- and three-dimensional simulations more meaningful, a 2D model is configured to be directly comparable to the 3D model. The 2D model used in these comparisons has the same domain length and depth and grid resolutions as the 3D model, and also uses warm rain microphysics only. We denote this as the coarse 2D model configuration. These two- and three-dimensional models are both used to simulate six cases with \(U_{\text{max}} = 15 \text{ m s}^{-1}\) and \(\sigma\) values of 0.75, 1.125, 1.5, 2.25, 3, and 4.5 km, producing corresponding values

![Fig. 10. Normalized area of instability vs lower-stratospheric Brunt–Väisälä frequency \((N_s)\) at 120 min. All cases use \(U_{\text{max}} = 15 \text{ m s}^{-1}\) and \(\sigma = 3 \text{ km}\). Squares show area from CT-CT500; crossed squares show area from CT500–25km.](image-url)
of $U_{\text{max}}/\sigma$ of 20, 13.33, 10, 6.66, 5, and 3.33 m s$^{-1}$ km$^{-1}$, respectively.

Figure 11 shows a comparison between the simulated cloud and wave breaking at 60 min for the (coarse) 2D model and a cross section through the 3D model at $y = 25$ km (the midpoint of the domain in the $y$ direction), for conditions corresponding to the baseline simulation. Although there are a number of differences in the details of the presented variables, there are many qualitative similarities. Both the 2D and 3D simulated clouds show similar depths (mostly controlled by the tropopause height), the altitude of the breaking approximately coincides, and it is those waves propagating downshear that break. However, the 2D model shows a wider cloud and greater area of cloudy air in the plotted domain in comparison with the 3D model.
and at this time the 3D cloud appears to be at an earlier stage in its life cycle with overshooting updrafts evident. The 2D and 3D simulations show gravity waves with similar horizontal scales, but as expected, the 2D model shows larger amplitude wave perturbations than the 3D model. Nonlinear effects will cause larger amplitude gravity waves to break more readily than smaller amplitude waves. This process is reflected in a comparison of the regions of nonzero \(K_M\) between Figs. 11a and 11b. The 2D model, with its larger amplitude waves, induces breaking closer to the cloud and the region of instability is substantially larger. Although there is less breaking in 3D, the 3D simulation does show a wave that is marginally unstable that appears close to breaking from the cloud top to about 17 km (\(x = 50\) km), which suggests that this wave might break throughout this depth at later times or if it had larger amplitude. The other main difference between the 2D and 3D simulations is that in the 2D model two phases of the gravity waves appear to be breaking, whereas in the 3D model only one phase is breaking above the cloud. This difference may simply be due to the two modeled clouds being at slightly different stages of their convective life cycle, with the wider 2D cloud featuring more complete wavelengths above cloud top.

Figure 12 compares the normalized areas of instability of the coarse 2D and 3D simulations, using a background wind profile with \(U_{\text{max}} = 15\) m s\(^{-1}\), and the six different values of the jet width parameter, \(\sigma\). The areas for the coarse 2D simulations are calculated in the same way as in section 2. The areas for the 3D simulations are calculated from cross sections along the \(x\) axis, with the result shown in Fig. 12 being the result of the average of the areas defined by five cross sections through \(y = 24.5, 24.75, 25.0, 25.25,\) and \(25.5\) km, surrounding the model domain centerline at \(y = 25\) km. This average is designed to reduce the influence of the 3D variability on the results.

In Fig. 12, the coarse 2D results show a similar trend to the 3D results, with both showing the largest normalized areas of instability at intermediate wind shear values. However, the optimum value is for a larger value of \(\sigma\) (3 km), that is, smaller background shear in 2D compared to 3D (\(\sigma = 1.5\) km). At their optimum value, the coarse 2D results show a normalized area of instability about 60% larger than the 3D results. The differences between the 2D and 3D results can be attributed to the different wave amplitudes. For example, at 15 km and \(\sigma = 3\) km the standard deviation (calculated over the entire 100-km length of the domain) of the \(x\) component of the horizontal velocity is 2.69 m s\(^{-1}\) for the 2D model and 1.04 m s\(^{-1}\) for the 3D model (along \(y = 25\) km); the 2D model has wave amplitudes about 2.5 times larger than the 3D model. Larger amplitude waves will break down more readily because of nonlinear effects and wave-induced critical levels (e.g., Breeding 1971; Fritts 1982). Specifically, if the wind is written as \(u = U + u'\), where \(U\) is the background wind and \(u'\) is the wave perturbation, a wave-induced critical level will occur if \((U + u') - c = 0\) (i.e., the larger the wave-induced amplitude \(u'\), the smaller the change in \(U\) required to induce wave breaking). Therefore, because of the larger amplitudes in 2D, less wind shear is required to induce wave breaking than in 3D, causing the optimum shear value to be at smaller shears (i.e., larger \(\sigma\) values) in 2D.

The results shown in Fig. 12 can be directly compared to the higher-resolution 2D results shown in Fig. 7a. The higher resolution results produce very similar normalized areas of instability to the coarse 2D results, and the optimum value of the wind shear coincides. This suggests that the results are not particularly sensitive to model resolution or microphysics parameterization.

In general, 3D models should produce more realistic representations of the occurrence of turbulence above convection because of gravity wave breaking and instabilities in the vicinity of the cloud boundary. However, as is the case here, computational limitations still restrict the size and resolution of model domains and the number of simulations that can be completed. These limitations result in coarser resolution and relatively smaller 3D domains in comparison with the 2D models.
Nonetheless, this brief comparison between the two- and three-dimensional results has shown that the relationships between turbulence area and wind shear produced by a 2D model are qualitatively comparable to a 3D simulation. The most important difference is that the 2D simulations seem to maximize the areas of instability at lower shear values in comparison to 3D. With the limited number of simulations available, it is not clear what other generalizations can be made. However, it should be remembered that the 2D results from section 2 are derived from a higher-resolution model, are within a larger model domain, and use more sophisticated microphysics than the (currently affordable) three-dimensional results. Therefore, it could be that, in fact, the higher-resolution 2D results represent the more robust set of solutions.

5. Discussion

The two-dimensional simulation results presented in section 3 illustrate that the occurrence and intensity of above-cloud gravity wave breaking and the resultant turbulence are sensitive to the time relative to convective initiation, the wind shear, and the stratospheric static stability. Most of the sensitivity that was determined is relatively intuitive. However, when the wave breaking was quantified in terms of the area of instability there was indication of an optimum value of the wind shear. Yet, when the wave breaking was quantified in terms of the integrated TKE, there was no obvious optimum value and the integrated TKE generally increased with increasing wind shear. To explain this, consider the schematic shown in Fig. 1, which represents intermediate wind shear conditions. At low values of the wind shear, the change in wind speed with height is not rapid, which results in the critical level occurring at higher altitudes; in fact in the limit of very small wind shear, the magnitude of the wind speed change may not be large enough to induce a critical level in the lower stratosphere at all. Thus there will be a reduction in wave breaking as the wind shear decreases. At stronger wind shears, however, the change in wind speed with height is rapid, which will result in the altitude of the critical layer being reduced from that depicted in Fig. 1. As the height of the critical layer is reduced to be closer to the cloud top, there is less available area (or volume in 3D) for the breaking to occur between the wave source and the critical layer. Therefore for higher shears the area of breaking becomes smaller because it must occur closer to the cloud top. The modeling results showed that this breaking in strong shear conditions is more intense, and therefore larger wind shear results in more intense and more localized wave breaking. At the largest wind shears, the height of the critical level is very close to the cloud top. In these cases it is hypothesized that the relatively small distance between the critical layer and the source of the waves (i.e., the cloud top) should influence the dynamics of the source in some way, producing a highly nonlinear response. This is a topic of continuing research.

As mentioned in section 3, changes in upper-tropospheric–lower-stratospheric (UTLS) wind shear and stability influence the cloud-top dynamics, including cloud width, updraft overshoot, gravity wave generation, and subsequent turbulence extent and intensity. We have made every effort to reduce the influence of these changes and incorporate them into our analysis. In particular, the use of the same low-level wind profile and initialization method for all cases produces convective systems of similar scale, intensity, and level of organization. Also, the influence of shear and jet strength on the cloud width was (at least partially) removed by examining the turbulence areas divided by this width. However, one effect that has not been discussed is the change in gravity wave spectrum as a result of changes in the wind shear in the UTLS. Beres et al. (2002) showed that variations in upper-tropospheric shear could modify the gravity wave momentum flux spectrum in the stratosphere. This effect remains an active area of research, and is difficult to quantify unambiguously. These modifications to the cloud and hence the wave source could only be removed through artificial specification of a fixed wave source. In our opinion, this simplification is too unrealistic for these applications because it does not incorporate the important coupling between the cloud and its environment.

To obtain an indication of the changes in the strength of the wave source relative to the convective intensity, the upper-tropospheric convective activity is quantified by calculating the spatial and temporal average of the square of the in-cloud vertical velocity, \( \langle w^2 \rangle \). The average is constructed at 10 km at every grid point that is within cloud, every 2 min from 60 to 120 min for each of the 2D cases that utilize the BIS sounding, and \( U_{max} = 15 \text{ m s}^{-1} \) (cases 3 and 5–12, Table 1). These averages are shown in Fig. 13a, and there is clear intercase variation that should influence the strength of the gravity waves generated. To investigate the influence of these changes on the above-cloud turbulence, the areas normalized by the cloud width for the layer CT500–25km (Fig. 5b) are further normalized by this convective activity and shown in Fig. 13b. This additional normalization does reduce the trends apparent in Fig. 7b, but the largest normalized areas still occur at intermediate shears. While this latter normalization does indicate some influence of changes in the wave source on the
turbulence generation, these effects are arguably irrelevant for the aviation application.

6. Summary

This study has examined the sensitivity of turbulence generation above deep convection to changes in background flow conditions, in particular the above-cloud wind shear and static stability. This turbulence can be due to shearing instabilities caused by the strong flow deformation near the cloud boundary, and also due to breaking gravity waves. The above-cloud turbulence was quantified in terms of the areas indicated by regions of low Richardson number, as well as the intensity of subgrid-scale turbulent kinetic energy. Understanding these sensitivities is important for the improved understanding of wave breakdown, dissipation, and propagation above deep convection, the generation of turbulence and mixing, and aviation safety.

A high-resolution two-dimensional cloud-resolving model was configured to produce deep and intense convection, and the background wind and static stability were systematically varied in a total of twenty-four simulations. These simulations represent the first such study that examines the sensitivity of aviation-scale turbulence above convection to systematic variations in background flow conditions. The simulations showed that the areas and intensities of wave breaking were highly sensitive to both the background wind shear and static stability. Weaker lower-stratospheric static stability produced a larger region of wave breaking than more stable conditions. The intensity of the wave breaking generally increased with the strength of the above-cloud wind shear; however the area of instability showed the largest values at intermediate shears. All of the simulations, including those containing very weak (or zero) wind shear, featured nonzero areas of mixing due to instabilities at the cloud boundary and nonlinear effects due to large amplitude gravity waves.

To assess the potential relevance of these two-dimensional simulations when applied to the real (three-dimensional) atmosphere, a number of coarser-resolution three-dimensional simulations were completed as well as comparable (coarse) two-dimensional simulations. These comparisons showed that the two-dimensional model produced qualitatively similar areas of instability and similar relationships with wind shear in comparison with the three-dimensional model, in that both show a maximum area of wave breaking and turbulence at intermediate values of the jet width parameter $\sigma$. The two-dimensional model did, however, show this optimum value of $\sigma$ to be smaller in magnitude.
(i.e., the shear is larger for the same $U_{\text{max}}$) in 3D compared to 2D. This difference was attributed to larger amplitude waves in the 2D simulations, which break down more readily than in the 3D simulations with smaller amplitude waves. This was the main qualitative difference between the 2D and 3D results; however, other differences in details remain, and these differences should be taken into account when interpreting the two-dimensional model results reported in this paper.

Both the 2D and 3D model simulations agreed with previous studies (Lane et al. 2003; Lane and Sharman 2006) that have shown that small-scale gravity waves produced by deep convection can readily interact with a critical level induced by a change in the above-cloud wind speed (i.e., wind shear). And, as in these previous studies, the simulations show that the areas of wave breakdown are contained in a fairly deep layer in the downshear regions above the cloud as waves of different wavelengths encounter levels close to their unique critical level. However, the major new result here is that the area of above-cloud turbulence tends to maximize at intermediate values of the above-cloud wind shear. Thus, the strength of the wind shear is of paramount importance in controlling the occurrence and intensity of breaking, regardless of the nominal strength of the background wind. This result has practical importance because the current FAA guidelines used by the aviation industry to avoid turbulence above convection are based solely on the cloud-top wind speed and not the wind shear. The results presented in this study illustrate that these guidelines should also include the effects of wind shear and static stability above cloud top; the appropriate formulation of such a guideline is the subject of continuing research.

The simulations presented here used a single functional form of the wind profile that features a jet structure with positive shear in the upper troposphere and negative shear in the lower stratosphere. Thus this case represents deep convective development below an upper-tropospheric jet, as was the case studied by Lane et al. (2003), which produces convection that can be classed as a leading stratiform system (LS; Parker and Johnson 2000). The environments that produce LS systems arguably present the most likely collocation of intense convection and strong environmental wind shear in the lower stratosphere. However, these LS systems are not the most common or long-lived convective systems (Parker and Johnson 2000), with trailing stratiform systems occurring more often and lasting longer. The results presented here can be applied to short-lived convective systems with strong upper-level background wind shear, which will control the wave breaking and turbulence. Long-lived convective systems, however, create their own circulation, which generates above-cloud wind shear with strength that cannot be determined in isolation from the convective system. This poses additional difficulties in utilizing these results for turbulence prediction or avoidance because it would be the cloud-induced circulations that would define or at least modify the above-cloud wind shear that causes the wave breaking rather than the environmental wind shear. Therefore, extending these results to long-lived systems should be a topic of future research.

This study has explored some of the sensitivities of above-cloud turbulence to background wind shear and lower-stratospheric static stability using one simple wind profile. Other sensitivities, including directional wind shear, cloud depth and intensity, the height of the jet in comparison with the tropopause, and midlevel shear, have not been examined and would undoubtedly show large sensitivities also. Therefore, the results presented here are necessarily limited in their scope and generality. Further sensitivity studies are required to better understand the processes controlling turbulence above and around convective clouds. Nevertheless, this study has clearly shown that current FAA guidelines for thunderstorm avoidance are oversimplified and lack a rigorous physical basis, and future research will focus on improving these guidelines.

Acknowledgments. This study was funded in part by the FAA Aviation Weather Research Program, by NASA CAN Contract NN06AA61A, by a University of Melbourne Early Carer Research Grant, and by an Australian Research Council Discovery Project (DP0770381). The views expressed are those of the authors and do not necessarily represent the official policy or position of the FAA. Thank yous are given to Drs. Terry Clark and Bill Hall for their assistance with the numerical model, to Dr. Stan Trier for his review of a draft version of this manuscript, and to the three anonymous reviewers for their useful comments.

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