Convection-Permitting Simulations of the Environment Supporting Widespread Turbulence within the Upper-Level Outflow of a Mesoscale Convective System

STANLEY B. TRIER AND ROBERT D. SHARMAN
National Center for Atmospheric Research, * Boulder, Colorado

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ABSTRACT

Widespread moderate turbulence was recorded on three specially equipped commercial airline flights over northern Kansas near the northern edge of the extensive cirrus anvil of a nocturnal mesoscale convective system (MCS) on 17 June 2005. A noteworthy aspect of the turbulence was its location several hundred kilometers from the active deep convection (i.e., large reflectivity) regions of the MCS. Herein, the MCS life cycle and the turbulence environment in its upper-level outflow are studied using Rapid Update Cycle (RUC) analyses and cloud-permitting simulations with the Weather Research and Forecast Model (WRF). It is demonstrated that strong vertical shear beneath the MCS outflow jet is critical to providing an environment that could support dynamic (e.g., shearing type) instabilities conducive to turbulence. Comparison of a control simulation to one in which the temperature tendency due to latent heating was eliminated indicates that strong vertical shear and corresponding reductions in the local Richardson number (Ri) to \( \leq 0.25 \) at the northern edge of the anvil were almost entirely a consequence of the MCS-induced westerly outflow jet. The large vertical shear is found to decrease Ri both directly, and by contributing to reductions in static stability near the northern anvil edge through differential advection of (equivalent) potential temperature gradients, which are in turn influenced by adiabatic cooling associated with the mesoscale updraft located upstream within the anvil. On the south side of the MCS, the vertical shear associated with easterly outflow was significantly offset by environmental westerly shear, which resulted in larger Ri and less widespread model turbulent kinetic energy (TKE) than at the northern anvil edge.

1. Introduction

Moderate-to-severe turbulence associated with thunderstorms constitutes a significant hazard to commercial aviation. Based on correlation of pilot reports of turbulence (PIREPS) and lightning flash data Wolff and Sharman (2008) concluded that \(~20\%\) of upper-level (5.5–18.3 km MSL) turbulence is thunderstorm related. Such estimates would likely be significantly greater if convection-induced turbulence (CIT) cases in which the turbulence occurs at significant distances (e.g., tens of kilometers or more) from active thunderstorms could be accounted for. These remote CIT events are of particular concern since they are typically more difficult to discern and avoid using standard hazard identification technologies (e.g., radar and satellite imagery, and lightning detection networks).

Remote CIT events at nominal jet cruising altitudes (e.g., 9–12 km MSL) can occur in a variety of locations relative to clouds and through several different mechanisms. For example, Lane et al. (2003) and Lane and Sharman (2008) simulate and discuss the role of vertically propagating gravity waves in the onset of CIT above rapidly growing thunderstorm updrafts that reach the statically stable tropopause. In contrast, Fovell et al. (2007) illustrated how circulations originating within the anvil region of a severe thunderstorm can reduce the Richardson number, \( \text{Ri} = N^2/\|\partial V/\partial z\|^2 \), to values that support turbulence production. There, the reduction of Ri relative to that of the surrounding environment for distances of \(~10\) km outside of cloud was accomplished by both increases in the vertical shear

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

Corresponding author address: Stanley B. Trier, National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307-3000.

E-mail: trier@ucar.edu

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magnitude $|\partial V/\partial z|$ and decreases in static stability $N^2 = (g/\theta)(\partial \theta/\partial z)$.

In the current study we examine how CIT can occur over more widespread areas and up to several hundred kilometers from active deep convection through similar reductions of $Ri$ from environmental values in the upper-level outflow region of a mesoscale convective system (MCS). MCSs (e.g., Maddox 1983) are a common mode of convective organization resulting in much of the heavy warm season rainfall over the central Great Plains and Midwestern United States (Fritsch et al. 1986). In a composite study of 10 MCS cases Fritsch and Maddox (1981) illustrated how outflow near the tropopause can result in an enhanced jet north and northeast of the MCS rainfall centroid with diminished winds to its south and southwest. They explained that Coriolis accelerations significantly modified the outflow and that the observation of strongest winds within the northeast quadrant of long-lived MCSs was related to the similarity in direction of the outflow and environmental winds. In contrast, weaker winds typically occur on the south and southwest flanks of the MCS, where the environmental flow opposes the storm-induced outflow.

Figure 1a illustrates the diurnally averaged association of heavy rain cases over the Great Plains during the 2003–06 warm seasons with 200-hPa zonal winds from Rapid Update Cycle (RUC; Benjamin et al. 2004) analyses. Here, the heavy rainfall maximum occurs at ~0800 UTC (slightly after local midnight) and precedes by several hours the RUC wind maximum, which occurs several hundred kilometers (~5° of latitude) to the north. The enhancement of the jet winds following the heaviest rainfall suggests strong influences on the upper-level flow by organized convection as described by Fritsch and Maddox (1981).

Weak or negative absolute vertical vorticity, with associated strong lateral gradients, occurs along and to the south of the MCS-induced anticyclonically curved outflow jet. Previous work (e.g., Knox 1997) has identified mesosynoptic regions with these flow properties as favored locations for clear-air turbulence (CAT). There, the resulting CAT may be associated with gravity wave emission resulting from inertial instability (Knox 1997) or other manifestations of the larger-scale flow imbalance (Knox et al. 2008). Other studies (e.g., Ellrod and Knapp 1992) have discussed vertical shear enhancements within the outflow being associated with turbulence within the northeast quadrant of an MCS.

In the current study, Fig. 1b confirms a significant increase in the average 200–400-hPa vector wind difference magnitude of 7–13 m s$^{-1}$, which approximately coincides with the location of the enhancement of the 200-hPa zonal jet (Fig. 1a) north of the rainfall maximum. This wind difference may be considered an estimate of the bulk vertical shear centered at ~9.5 km MSL (~32 000 ft), which is a common cruising altitude for commercial aircraft. Both the magnitude and temporal enhancement of the vertical shear north of the

![Fig. 1. Spatiotemporal relationships of warm-season heavy rainfall (color scale) to RUC analysis (a) 200-hPa zonal winds (contour interval of 1 m s$^{-1}$) and (b) 200–400-hPa vector wind difference magnitude (0.5 m s$^{-1}$ contour interval) for the heaviest 25% of rainfall cases in a 4-yr warm season climatology. Here, the rainfall cases are individually shifted so that their longitudinally averaged (90°–100°W) maxima for each of the composited events occur at 40°N latitude with RUC fields averaged over the same longitude belt shifted the same latitudinal distance. The rainfall is radar derived using the WSI Corporation NOWrad national composite of the Weather Surveillance Radar-1988 Doppler (WSR-88D) network at 15-min frequency and applying the reflectivity–rain rate relationship ($Z = 300R^{1.5}$), where $Z$ is the reflectivity factor (mm$^6$ m$^{-3}$) and $R$ is the rainfall rate (mm h$^{-1}$).]
heavy rainfall are likely much greater in individual cases. This is because the MCS-outflow jet and the associated vertical shear beneath its core typically occupy layers of only a few kilometers, whose altitude varies a comparable depth among the cases averaged in the climatological composite (Fig. 1b).

In addition to the vertical shear increases documented in Fig. 1b, the thermal stratification in ice-saturated outflows (e.g., MCS anvils) near the tropopause is much weaker than at lower levels. Together, these factors suggest that regions near the outer edges of MCS upper-level outflows may be particularly susceptible to CIT at commercial aircraft flight levels. While MCSs are primarily a nocturnal phenomenon, Fig. 1 also suggests that enhanced upper-level outflow and by implication, the associated turbulence threat, often persists for several hours beyond sunrise when commercial air traffic is increasing.

In this study we analyze convection-permitting simulations of an observed case of widespread CIT reported along the northern and northeastern edges of a large nocturnal MCS cloud shield. The goal is to better understand the relationship between the MCS and CIT. Section 2 presents flight-level observations of the turbulence, radar observations of the MCS life cycle, and diagnoses of its large-scale environment based on hourly RUC analyses. Section 3 describes the numerical model and our design of the numerical simulations. An important aspect of our modeling approach is the comparison of a control run that allows grid-resolved deep convection to an otherwise identical “adiabatic” run, in which the temperature tendency due to latent heat release is neglected. In this way, the effects of deep convection on the mesoscale environment that may support turbulence generation can be directly assessed. In sections 4 and 5 we illustrate how vertical shear and thermodynamic stratification are significantly modified over distances of hundreds of kilometers in a manner that supports the generation of turbulence in specific regions of the MCS upper-level outflow from both static and dynamic (e.g., shearing) instabilities.

2. Observations of the 17 June 2005 central U.S. turbulence case

a. Flight-level in situ data

Three eastbound Boeing 757 commercial airline flights equipped with automated turbulence detection systems (Corman et al. 1995) based on atmospheric eddy dissipation rate \( \epsilon^{1/3} \) encountered light-to-marginally severe turbulence \( \left( \epsilon^{1/3} = 0.10–0.40 \text{ m}^2/\text{s}^{3/2} \right) \) along a \( \sim 500\)-km path from western Kansas into north-central Missouri from 0700 to 1000 UTC 17 June 2005. This widespread turbulence event occurred near the northern edge of an extensive MCS upper-level outflow whose cirrus cloud shield is depicted in Fig. 2. There were also PIREPS of moderate turbulence spanning a region from southeast Colorado to the southeast Iowa–northeast Missouri border area from five additional flights that traversed the MCS cloud shield during 0300–1200 UTC 17 June. An interesting aspect of the turbulence encounters on the three instrumented aircraft (Fig. 2) is that while occurring within the upper-level outflow, they were each located several hundreds of kilometers north and east of the active deep convection regions (as indicated by large radar reflectivity) within the mature MCS (Fig. 3b).

Nearly continuous moderate turbulence \( \left( \epsilon^{1/3} > 0.2 \text{ m}^2/\text{s}^{3/2} \right) \) with maximum values that were marginally severe \( \left( \epsilon^{1/3} \sim 0.4 \text{ m}^2/\text{s}^{3/2} \right) \) was recorded by flight 2 between pressure altitudes of 11.3 and 11.6 km (all heights in this subsection are pressure altitudes)\(^{1}\) for \( \sim 160\) km in northeast Kansas from 0744 to 0755 UTC (Fig. 2). This aircraft descended over northwest Missouri where mostly lighter turbulence was recorded below 11 km and was avoided altogether beneath 10 km. An earlier flight (flight 1) with a similar but more northern flight path (Fig. 2) also recorded light turbulence over western Kansas but apparently avoided the more intense turbulence by descending to 10.1 km over extreme north central Kansas. The third flight followed the Los Angeles to New York route of flight 2, but \( \sim 2\) h later, similarly recorded light turbulence over western Kansas from 0920 to 0943 UTC. As with flight 2, the third flight recorded moderate turbulence over northeast Kansas and northwest Missouri (though more intermittent than flight 2) from 0944 to 0957 UTC between 11.3 and 11 km before descending farther.

b. Overview of the 16–17 June MCS and the characteristics of its inflow environment

By late evening the MCS comprised two prominent convective bands. One was located along its south-southeastern edge and the other was along its northwest periphery (Fig. 3a). The northwest band was relatively short lived \( (T \sim 3\ h) \). However, simulations discussed in sections 4 and 5 suggest that upper-level outflow from this MCS component was an important contributor to the widespread turbulence several hours later.

\(^{1}\) Calculations using the hypsometric equation (e.g., Holton 1992, p. 20) with RUC soundings and surface pressure data indicate that the geometric heights of the recorded turbulence for all flights are probably 0.25–0.3 km greater than the pressure altitudes, which are based on the U.S. Standard Atmosphere, 1976.
An RUC analysis sounding (Fig. 4a) at S1 in the vicinity of the northwest band (Fig. 3a) indicates that intense deep convection was supported by conditionally unstable layers that surmounted a shallow surface-based stable layer. The upper layer from 740 to 660 hPa had convective available potential energy (CAPE) ranging from 600 to 1600 J kg$^{-1}$ with negligible convective inhibition (CIN) because of its nearly dry adiabatic structure. The most unstable air with average CAPE of 2950 J kg$^{-1}$ was located in the 870–760-hPa layer and was associated with a southwesterly low-level jet (LLJ). Although this air had moderate average CIN of 115 J kg$^{-1}$, it likely participated in deep convection once the convection became widespread.

The environment of the southeast convective band at S2 had deep surface-based moisture (Fig. 4b) resulting in average CAPE of 3650 J kg$^{-1}$ in the lowest 70 hPa. CIN was <70 J kg$^{-1}$ in this layer and negligible within the remnants of a daytime well-mixed layer directly above (890–840 hPa). The southeast band’s reorientation normal to the low-level shear (cf. Figs. 3b and 4b) together with only small overnight equivalent potential temperature $\theta_e$ reductions in its LLJ inflow likely influenced its greater longevity than the northwest band and contributed to the overall southward component of MCS propagation (Fig. 5).

Convection that was not directly associated with the MCS occurred closer to the reported turbulence (Fig. 3). This convection, which was weaker, and less persistent than the MCS convection, may have locally influenced the turbulence (as discussed in section 5). However, it cannot explain the widespread nature of the turbulence.

An RUC analysis sounding (Fig. 4c) at S3 (Fig. 3a) shows modest average CAPE of 900 J kg$^{-1}$ for air with only 20 J kg$^{-1}$ of CIN in the elevated layer from 890 to 840 hPa with an equilibrium level (EL) beneath 275 hPa ($Z < 10$ km). The relatively shallow EL is consistent with the maximum height of simulated storm tops in this region being several kilometers below the levels of the simulated MCS outflow and reported turbulence.

c. RUC analysis of the large-scale environment near flight level

A time–latitude diagram (Fig. 5) of the RUC analyzed 200–300-hPa wind vector difference magnitude reveals vertical shear enhancement several hours after and several hundred kilometers north of the heavy rainfall. These spatiotemporal relationships between the heavy rainfall and vertical shear strongly resemble those from the warm-season climatology of heavy rain cases (Fig. 1b).
Horizontal plots of 12.5-km MSL RUC-analysis winds (Fig. 6) confirm that enhancement of the longitudinally averaged vertical shear near 40°N (Fig. 5) is associated with the development of anticyclonic outflow, which is well established by 0800 UTC 17 June (Fig. 6b). At this time an extensive region susceptible to turbulence (i.e., Ri < 1), which was not present during the afternoon (Fig. 6a), occurred beneath the anticyclonic jet (Fig. 6b).

North–south-oriented vertical cross sections (Fig. 7) averaged for 400 km in the zonal direction (the approximate flight path direction in Fig. 2) indicate that the zone of small Ri was ~1 km deep and that the aircraft were flying near its northern edge. Both strong vertical shear (Figs. 7b,c) and weak static stability (Fig. 7a) contributed to the small Ri. Vertical shear zones occurred both above and beneath the jet, but small Ri was most extensive near flight levels below the jet because of the lesser static stability (Fig. 7a).

3. Numerical model and experimental design

a. Atmospheric model

The numerical model used for the high-resolution simulations is the Advanced Research Weather Research and Forecasting (ARW-WRF version 2.2) Model (Skamarock et al. 2005). The model was integrated for 14 h from 2100 UTC 16 June 2005 to 1100 UTC 17 June 2005 over a single subcontinental domain (shown later in Fig. 11). The model contains 600 × 500 horizontal
The vertical grid has 64 levels with spacings of ~60–170 m in the lowest 1 km, ~400 m between 1 and 15 km, and ~500–1500 m from 15 km to the model top near 31 km. A diffusive damping applied in the top ~10 km of the model strongly reduces the amplitude of vertically propagating waves and thus mitigates undesirable effects of their reflection off the rigid upper boundary. The initial and lateral boundary conditions are obtained from a synthesis of 1) hourly RUC native hybrid coordinate analyses (surface to ~100 hPa) and 2) 1.0° by 1.0° grid points, with 3-km spacing. The vertical grid has 64 levels with spacings of ~60–170 m in the lowest 1 km, ~400 m between 1 and 15 km, and ~500–1500 m from 15 km to the model top near 31 km.

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6-hourly National Centers for Environmental Prediction (NCEP) Global Final (FNL) pressure-level analyses (from \( p < 100 \) to \( \sim 10 \) hPa) linearly interpolated to 1-h temporal resolution.

b. Physical parameterizations

The 3-km horizontal grid spacing, while not ideal for completely resolving individual convective cells, explicitly resolves salient mesoscale aspects of the deep convection (e.g., Weisman et al. 1997; Bryan et al. 2003), which obviates the need for cumulus parameterization. The simulations discussed in the paper use the Thompson et al. (2008) bulk microphysical parameterization, which predicts cloud water, cloud ice, rain, snow, and graupel hydrometeor species. This particular scheme was selected because of its increased sophistication in the specification of the snow size distribution, shape, and density parameters over other single-moment schemes currently available in the WRF modeling system. This aspect of the scheme is potentially relevant considering our focus on the turbulence environment near and within the ice anvil regions of MCSs. Other necessary physical parameterizations were obtained using the Rapid Radiative Transfer Model (RRTM) longwave (Mlawer et al. 1997) and Dudhia (1989) shortwave radiation schemes, and the Noah land surface model (Ek et al. 2003).

The selected PBL scheme (Janjić 1990, 1994) predicts turbulent kinetic energy (TKE) and governs the vertical mixing between individual layers both within the PBL and at other levels in the model. At this writing, there is no advection of TKE in the ARW-WRF version of this scheme (J. Dudhia 2008, personal communication). Here, local forcing of TKE is provided by shear production \(- \overline{wU'}(\partial U'/\partial z) - \overline{wV'}(\partial V'/\partial z)\), buoyancy production \(\beta \overline{w\theta_L'}\), and dissipation terms, where \(\theta_L\) is the liquid water virtual potential temperature (e.g., Emanuel 1994), \(\beta\) is an empirical constant, and other variables have their conventional meanings. In the shear and buoyancy production terms, the lower case variables compose the subgrid fluxes, which are parameterized in terms of resolved-scale variables with \(- \overline{wU'} = K_M(\partial U'/\partial z)\), \(- \overline{wV'} = K_M(\partial V'/\partial z)\), and \(- \overline{w\theta_L'} = K_H(\partial \theta_L'/\partial z)\), where \(K_M\) and \(K_H\) are vertical exchange coefficients for momentum and heat, respectively. The horizontal mixing in the model is determined using a Smagorinsky first-order closure detailed in section 4.1.3 of Skamarock et al. (2005).

c. Simulations

The control simulation utilizing the above physical parameterizations is the primary focus of the forthcoming analysis. However, since we are interested in the effects of MCS-generated outflow on the environment of turbulence, we compare the control simulation with an additional simulation, in which the effects of buoyant deep convection are eliminated. This model run, which we term the “adiabatic simulation,” is configured identically to the control simulation except that the temperature tendency from latent heating/cooling processes in the microphysical parameterization is set to zero at all time steps.2 We also analyzed a third simulation that used an alternative bulk microphysical parameterization (Hong et al. 2004) but was otherwise identical to the control simulation. Apart from minor differences in MCS precipitation and TKE, the results from this third simulation are similar to the control simulation and will not be discussed.

4. The MCS upper-level outflow

a. Overview of the simulated MCS

Overall, the control simulation (Fig. 8) provides a reasonable representation of the observed MCS (cf. Fig. 3). As in other MCS simulations with horizontal resolutions that are marginally sufficient for resolving deep convection (e.g., Trier et al. 2006), the convective bands are somewhat too broad and intense. However, the control simulation captures well the locations of the observed convective bands near their maximum intensity (Figs. 3a and 8a) and the subsequent weakening of the northwest convective band (Figs. 3b and 8b). As in the observations (Fig. 3), the simulated (Fig. 8) weaker stratiform precipitation is mostly north of the active convection.

Maximum simulated TKE near the aircraft flight levels is greatest during the intensification stages of the MCS (Fig. 8a), but is more widespread later in the MCS life cycle (Fig. 8b), closer to the time of the observed turbulence. In the remainder of this paper, we focus on factors that produce the simulated TKE during this mature and somewhat less intense stage of the MCS.

b. Effect of the MCS on the upper-level flow conditions

A large circular region of low potential vorticity (PV) at 12.5 km (where typically PV > 1.5 PVU) indicates widespread upward transport of lower-tropospheric air by the MCS circulation (Fig. 9a). The wind speed maximum in the northeast quadrant of the low PV region, with a wind speed minimum on its southwest flank, is consistent with the results of Fritsch and Maddox.

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2 This sensitivity experiment is not strictly adiabatic since it retains subgrid mixing and radiative processes.
Unlike the maximum outflow winds, the maximum vertical shear directly below their core is confined to a narrow latitude band and extends well westward from where the upper-level winds are strongest (Fig. 9a). The change of predominately southerly (195°) to westerly (250°) momentum during the ~3 h after trajectories exit the northwest convective band constitutes approximately 1/6 of the arc length of a circle. Since 1/6 of the inertial period at the MCS location is $T/6 = \pi/6\Omega \sin(38°) \approx 3.2$ h, the 3-h change in wind direction along trajectories is consistent with Coriolis accelerations.

However, Coriolis accelerations cannot account for the 5–10 m s$^{-1}$ increases in wind speed that occur when trajectories first enter the upper-level outflow between 0500 and 0600 UTC (Fig. 9d). Here, the northeastward flow acceleration is aided by a strong unbalanced horizontal pressure gradient force associated with the mesohigh (e.g., Fritsch and Brown 1982) situated above the level of neutral buoyancy near the region of strong convection (Fig. 10a).

Overall, our cloud-resolving simulations of an observed MCS resemble the simulations of Wolf and Johnson (1995) obtained with idealized initial conditions and parameterized deep convection in a global channel model. In that study, strong accelerations in the direction of the environmental horizontal pressure gradient force were similarly enhanced by deep convection, with the unbalanced flow undergoing significant inertial turning toward a more balanced state in the downstream jet streak. The transition toward a more balanced flow along the control simulation upper-level outflow of the current case is suggested by the much smaller wind-crossing angle with the isobars over eastern Missouri–western Illinois (Fig. 10c) than evident earlier closer to the outflow source over northwestern Oklahoma (Fig. 10a).

c. Comparisons between the control and adiabatic simulations

The broad region of reduced $\text{Ri} < 1$ below the outflow jet in the control simulation over northern Kansas and Missouri (Fig. 10c) is consistent with the 0800 UTC RUC analysis (Fig. 6b). The absence of small $\text{Ri}$ in the adiabatic simulation is related to much weaker upper-level flow (Fig. 10d). Here, the 12.5-km flow over northeastern Kansas is situated on the west side of a southward-moving trough (Figs. 10b,d). In contrast to the control simulation, this flow becomes more northerly and weaker after the 2100 UTC 16 June RUC analysis time (Fig. 6a).
The 12.5-km control–adiabatic simulation wind difference vectors (Fig. 11a) illustrate the large ($L > 1000$ km) horizontal scale of the MCS-induced outflow as noted in past studies of MCS upper-level outflows (e.g., Fritsch and Maddox 1981). The MCS-induced upper-level outflow is reflected in the control–adiabatic wind shear vectors from the 11- to 12.5-km layer directly below (Fig. 11b). The effects of the MCS-induced upper-level outflow (Fig. 11a) and its associated vertical shear (Fig. 11b) on the vertical shear in the control simulation is examined in Fig. 12 at different locations along the anvil cloud edge of the control simulation for a 90-min period (0700–0830 UTC) that comprises the observed turbulence.

The north (N) location (Fig. 11b) is situated toward the eastern edge of where moderate turbulence was observed (Fig. 2), and contains the strongest simulated vertical shear (Fig. 12). Toward the western edge of where moderate turbulence was observed (north-central Kansas), the vertical shear was of comparable magnitude but had a greater southerly component (not shown) due to the more southerly direction of the MCS outflow (Fig. 11a) and its associated vertical shear (Fig. 11b). In these locations the environmental (i.e., adiabatic) flow and its associated vertical shear are weak (e.g., Figs. 12a,b). Thus, the strong vertical shear in the control simulation results almost entirely from dominance of the outflow (control–adiabatic) and its associated vertical shear beneath jet level.

The strongest vertical shears near flight levels where the actual turbulence was observed (gray shading in Figs. 12a,b) are reflected in the vertical profile of time-averaged $R_i$ at N (Fig. 13), which indicates $R_i \sim 0.25$ between $Z = 11$–$12$ km. The larger $R_i$ value in the secondary minimum above the jet between $Z = 13$–$14$ km

Fig. 9. (a) 0750 UTC 17 Jun 2005 12.5-km Ertel potential vorticity <1.5 PVU (medium gray), 1.5–3 PVU (light gray) and >3 PVU (white), horizontal wind speed (thin solid contours, 10 m s$^{-1}$ contour intervals), and magnitude of the 11–12.5-km vector wind difference (bold dashed contours, 5 m s$^{-1}$ contour intervals) from the control simulation. Arrowheads indicate origin of 3-h back trajectories at 0750 UTC discussed in the text. (b) Zonal wind component, (c) meridional wind component, and (d) wind speed along trajectories.
is a consequence of greater static stability rather than weaker vertical shear.

Unlike at N, the environmental winds at the south (S) location of Fig. 11b strongly oppose the outflow (Figs. 12c,d). In particular, the environmental westerlies more than offset the outflow easterlies, resulting in relatively weak westerly flow at the outflow level in the control simulation. Furthermore, the environmental shear, although weaker than the outflow shear, significantly reduces the total vertical shear in the control simulation. As at N, maximum vertical shear beneath the outflow jet at S results in small Ri (Fig. 13). However, here, the smallest Ri values near 1 are significantly greater than those at N, where the vertical shear is much stronger.

5. Factors influencing simulated turbulence in the MCS upper-level outflow

In this section we examine how control simulation TKE near the flight levels of the observed turbulence is linked to processes occurring within the MCS upper-level outflow. Since the simulated TKE results from a
parameterization (section 3) in a relatively coarse model ($\Delta x = 3$ km, $\Delta z \approx 400$ m at flight levels), it does not necessarily represent the small scales of turbulence that the aircraft experience. The assumption inherent in the forthcoming analysis is that the spatial distribution of the simulated TKE provides some indication of where environmental conditions are conducive to turbulence generation at smaller scales.

a. Simulated time-averaged conditions during the period of observed turbulence

Expansive regions of TKE, large vertical shear, and small moist static stability at 11.75 km occur toward the edge of the anvil cloud on the north side of the MCS (Fig. 14) and are strongly interrelated. For the moist static stability we have used the formulation of the Brunt–Väisälä (buoyancy) frequency from Durran and Klemp [1982, their Eq. (36)] modified to account for ice (in addition to liquid) latent heating/cooling considerations. During the 90-min period of observed moderate turbulence from 0700 to 0830 UTC, time-averaged 11.75-km TKE is most widespread in this northern region several hundred kilometers away from the MCS convection (Fig. 14a). Elsewhere, localized patches of significant time-averaged TKE are associated with the steadiest regions of intense convection within the MCS. At this altitude, the greatest time-averaged vertical shear is near the north anvil edge (Fig. 14b) and approximately coincides with the widespread TKE region (Fig. 14a), whereas the smallest time-averaged moist static stabilities occur both farther south in the anvil and within the region of MCS convection (Fig. 14c).

Three distinct simulated cloud features are evident in corresponding time and zonally averaged north–south-oriented vertical cross sections (Fig. 15). These include widespread active deep convection ($x = 150$–350 km), its associated downstream anvil ($x = 350$–900 km), and shallower, less widespread convection underneath the anvil ($x = 600$–900 km).

The strength and shallowness of the upper-level outflow at large distances from the MCS convection explain why the vertical shear is strongest near the northern edge anvil edge (Fig. 15). Here, maximum TKE occurs near flight levels and is collocated with both strong zonal (Fig. 15b) and meridional (Fig. 15c) vertical shears near and underneath the shallow core of the outflow jet. Closer to the MCS convection ($x = 500$–600 km), the strongest vertical shear and TKE are located several kilometers lower at the interface between the upper-level outflow and a strong northwesterly rear-inflow jet (Fig. 15). Collander et al. (2006) similarly found strong vertical shears and the potential for turbulence near or within MCS-induced rear-inflow jets.

b. Temporal variability

Although the foregoing time averages reveal widespread TKE in the northern region of the MCS anvil, 3 Including ice effects results in a 5%–10% reduction in 11.75-km static stability relative to the corresponding dry static stability for area averages over the gray-shaded rectangular region in Fig. 14.
substantial temporal variability of simulated TKE and associated kinematic and thermodynamic variables occurs within this region during the MCS life cycle. This variability is apparent in the time series of $\theta_e$ near the core of the outflow jet at $z = 12.5$ km (Fig. 16). Figure 14 indicates the location of these examples, with point A (Fig. 16a) situated near the north anvil edge, and point B (Fig. 16b) located within the anvil ~250 km southwest of point A.

Spectral analyses of these and other time series (not shown) from locations within the gray-shaded rectangular region of Fig. 14 revealed a broad distribution of peaks with periods ranging from about 15 min to 3–4 h. The higher-frequency variations in both $\theta_e$ (Fig. 16) and the horizontal and vertical velocities (not shown) are likely associated with gravity waves, which are evident in animations using 5-min model output. Gravity wave sources apparent from these animations include the active MCS convection ($x = 150–350$; Fig. 15) and the weaker, less widespread convection farther north beneath the anvil ($x = 600–900$; Fig. 15). Both deep tropospheric heating (e.g., Pandya and Durran 1996) and
Fig. 14. Time-averaged (a) 11.75-km MSL turbulent kinetic energy (1 m² s⁻² contour interval starting at 1 m² s⁻²), (b) 11–12.5-km MSL wind difference magnitude (5 m s⁻¹ contour interval starting at 20 m s⁻¹), and (c) 11.75-km MSL moist static stability (1 × 10⁻⁵ s⁻² contour interval starting at 0 s⁻² and ending at 3 × 10⁻⁵ s⁻²) using 5-min model output from the control simulation from 0700 to 0830 UTC 17 Jun 2005 (t = 10–11.5 h). Maximum model-derived reflectivity in a vertical column at 0800 UTC is indicated by the color scale. The dashed south–north-oriented rectangle indicates the location of the zonally averaged south–north vertical cross section displayed in Fig. 15 and the gray shading indicates the region where the area-average and budget time series in Fig. 19 are calculated. The symbols A and B denote the location of the time series presented in Figs. 16 and 17.

Fig. 15. Time and zonally averaged south–north vertical cross sections for the region indicated by the dashed rectangle in Fig. 14 computed for the control simulation using 5-min model output from 0700 to 0830 UTC 17 Jun 2005 (t = 10–11.5 h) with fields of (a) potential temperature (black contours with 2-K intervals), (b) zonal wind component (black contours with 2.5 m s⁻¹ intervals), and (c) meridional wind component (black contours with 2.5 m s⁻¹ intervals, negative values dashed). In (a)–(c) the green contours are turbulent kinetic energy (0.5 m² s⁻² intervals), the red contours are the moist static stability <4 × 10⁻⁵ s⁻² (1 × 10⁻⁵ s⁻² intervals), and the gray/color shading is the total cloud condensate (scale at right). In (a) the thick solid rectangle indicates the horizontal location of the area-averaged calculations of Figs. 19a,b.
intense updraft cells overshooting their EL and encountering strong static stability at the tropopause (e.g., Lane et al. 2003) could generate gravity waves within the MCS convection region that propagate horizontally within the anvil outflow. Although the more isolated convection north of the MCS has updrafts that typically terminate a few kilometers beneath the anvil, this convection impinges upon a layer of enhanced tropospheric stability (Fig. 15a) and vertical shear (Figs. 15b,c) that could also support gravity wave propagation.

While high-frequency gravity waves are ubiquitous features of the simulation, the largest \( \theta_e \) perturbations (\( \sim 2–4 \) K) within the anvil are clearly associated with low-frequency variability (Fig. 16). This variability with periods of 3–4 h is tied to intensification and decay of mesoscale components of the MCS; most notably the northwest convective band (Fig. 8). Here, positive \( \theta_e \) perturbation development spanning features 1–2 in Fig. 16 follows the upstream intensification of the northwest band convection while negative \( \theta_e \) perturbation development spanning features 2–3 in Fig. 16 follows the decay of this convective band.

Since the dry and moist adiabatic lapse rates are nearly equal in upper-level (\( Z = 11–14 \) km) ice-saturated anvil clouds, the potential stability \( \partial \theta_e / \partial z \) is nearly proportional to the moist static stability. Large-amplitude low-frequency \( \theta_e \) perturbations occur throughout the anvil. The phase of these perturbations can vary substantially with height toward the downstream edge of the outflow due to effects of vertical shear (section 5c). Because of these phase differences with height, the low-frequency \( \theta_e \) perturbations within the 12.5-km outflow core at point A near the anvil edge (Fig. 16a) are associated with high-amplitude moist static stability variations at the 11.75-km approximate flight level below (Fig. 17a).

The vertical shear at point A is consistently strong (Fig. 17a), which provides an environment of low \( Ri < 1 \) (Fig. 17b). However, the low-frequency moist static stability variations are the primary cause of further lowering of \( Ri \) to values \( \sim 0.25 \), which are in turn associated with two 1–2-h episodes of enhanced TKE in this location (cf. Figs. 17a,b). High-frequency variations are evident in the vertical shear and moist static stability time series (Fig. 17a) and are likely associated with gravity waves. These high-frequency variations also impact \( Ri \) and TKE (Fig. 16b), but to a much lesser extent than the low-frequency variations.

c. Low-frequency thermodynamic destabilization in the anvil

Mesoscale processes contributing to the moist static stability decreases within the upper-level outflow are now examined for the episode that spans features 2–3 at locations A and B in the 12.5-km \( \theta_e \) time series (Fig. 16). Locations A and B are situated along a vertical cross section, averaged 150 km in the cross line direction (Fig. 18a, gray shading), approximately parallel to upper-level outflow trajectories emanating from the decaying northwest convective band (cf. Fig. 8). Back trajectories in Fig. 18a are calculated from the region of maximum geopotential height along the 345-K \( \theta_e \) surface starting at 0830 UTC.

Within this cross section the evolution of both the 342- and 345-K \( \theta_e \) surfaces within the upper-level anvil outflow are displayed (Figs. 18b,c). The vertical distance between these surfaces provides an indication of the potential stability. Because of its simple conservation relation, examining the evolution of the potential stability,

\[
\frac{\partial}{\partial t} \left( \frac{\partial \theta_e}{\partial z} \right) = -\frac{\partial}{\partial z} (\mathbf{V} \cdot \nabla \theta_e) - \frac{\partial}{\partial z} \left( w \frac{\partial \theta_e}{\partial z} \right),
\]

FIG. 16. Time series of 12.5-km MSL equivalent potential temperature perturbation from 6.5-h time mean at (a) point A and (b) point B whose locations are displayed in Figs. 14 and 18. The annotations 1, 2, and 3 highlight the approximate temporal occurrence of low-frequency extrema of the perturbations discussed in the text.
facilitates interpretation of physical processes that influence the moist static stability. In (1), \( u_e \) is the equivalent potential temperature in the ice anvil cloud with \( u_e \), \( L \), \( q_i \), \( c_p \), and \( T \) representing potential temperature, latent heat of fusion, saturation mixing ratio with respect to ice, specific heat at constant pressure, and temperature, respectively. We have neglected mixing and radiative effects in this form of the conservation equation. Thus, local potential stability changes \( \text{lhs of (1)} \) result only from differential advection by the horizontal wind \( \text{first term on rhs of (1)} \) and from effects of vertical motion \( \text{second term on rhs of (1)} \).

Figure 18b indicates the upward displacement of \( \theta_e \) surfaces in the region of the 0700–0830 UTC time-averaged mesoscale updraft, which itself is a feature commonly observed in the stratiform anvil region of mature-to-dissipating MCS (e.g., Gamache and Houze 1982; Rutledge et al. 1988; Biggerstaff and Houze 1991; Cifelli and Rutledge 1994; Knupp et al. 1998). The upward displacement of the \( \theta_e \) surfaces is associated with the development of a horizontal \( \theta_e \) gradient at the northeast edge of the mesoscale updraft, which is evident by 0745 UTC (Fig. 18b). The strong vertical shear beneath the outflow jet core leads to a further steepening of the \( \theta_e \) surfaces and thermodynamic destabilization well downstream of the mesoscale updraft (near \( x = 400 \) km) through differential horizontal advection by 0830 UTC (Figs. 18b,c).

Thus, in addition to the direct role strong vertical shear plays in lowering \( R_i \), it plays an important indirect role through influences on potential stability. These influences encourage turbulence production by both lowering the vertical shear threshold necessary to support dynamic instabilities (i.e., \( 0 \leq R_i \leq 0.25 \)) over broad regions and by generating localized regions of potential instability or moist static instability \( (R_i < 0) \) within the anvil.

Note that the \( \theta_e \) reduction phase at point B (Fig. 16b) near the northeast edge of the mesoscale updraft (Fig. 18b) precedes that at point A (Fig. 16a) by \( \approx 100 \) min, which is the approximate time period in which horizontal advection in the 40 m s\(^{-1}\) outflow jet (Fig. 18c) accounts for such a \( \approx 250\)-km displacement. Thus, while the upper-level outflow could be influenced by low-frequency gravity waves resulting from changes in upstream heat sources (e.g., Pandya and Durran 1996) including the northwest convective band (Fig. 8), the approximate advective motion of the \( \theta_e \) perturbations suggests a quasi-horizontal plumelike behavior. The precise mechanism underlying these low-frequency variations is clearly an important unresolved issue pertaining to the turbulence environment that warrants further study.

d. Area-averaged conditions and thermodynamic budget near the anvil edge

We now confirm both the widespread nature of the low-frequency variations and their link to mesoscale physical process by performing quantitative budgets for the shaded rectangular region in Fig. 14. In this region, which extends \( \approx 150 \) km southward from the north anvil cloud edge, the area-averaged TKE lags the area-averaged vertical shear (Fig. 19a). This phase shift is consistent with shear production of TKE (section 3), and the subsequent reduction of the shear by subgrid vertical mixing during periods of large TKE. The replenishment of vertical shear following its reduction is a consequence of continuous MCS-induced outflow emanating from upstream.

While the vertical shear and TKE are generally out of phase, area-averaged TKE increases from 0430 to 0530 UTC, and 0800 to 0900 UTC coincide with sharp area-averaged moist static stability decreases (Fig. 19a).
Under conditions of moist static stability, resolved-scale processes and subgrid vertical mixing can act in the same direction to reduce the moist static stability. However, the close agreement between the sum of the forcing terms and tendency term in a budget analysis of potential stability (Fig. 19b) based on (1) confirms the dominance of the resolved-scale processes on stability changes in the control simulation. Here, potential stability decreases in the northern anvil region are dominated by the differential horizontal advection term. However, as suggested by the cross-section analysis of section 5c, vertical motion (not shown) plays a significant role through upward displacement of the isentropes that contribute to the horizontal gradients at the southern edge of the budget calculation region (box within Fig. 15a), which in turn influence the differential horizontal advection over the region.

6. Summary and discussion

A convection-permitting configuration of the WRF model was used to simulate the environment of widespread moderate turbulence recorded by three commercial airline flights on 17 June 2005. Comparison of conventional radar products with RUC model analyses revealed that the turbulence occurred within a mesoscale region of small Richardson number (Ri < 1) situated beneath an asymmetric anticyclonic outflow jet,
several hundred kilometers north of the deep convection within a nocturnal MCS. The WRF helped to elucidate MCS-induced modifications of the environment that may have accounted for the turbulence by 1) affording a comparison between a control simulation with grid-resolved deep convection and an "adiabatic" simulation in which the effects of buoyant deep convection were not permitted, and by 2) providing a high spatial and temporal resolution model dataset from the control simulation that facilitated diagnosis of mechanisms responsible for the evolution of the turbulence environment.

The control simulation captured the nocturnal MCS organization, which consisted of two primary convective bands and a stratiform anvil region located to their north. Trajectory analysis indicated the portion of the upper-level outflow that was associated with underlying turbulence near the northern anvil edge emanated from the northwest flank of the MCS convection. When first exiting the deep convection, this outflow was southerly and experienced significant speed increases due to an unbalanced horizontal pressure gradient force associated with the MCS upper-level mesohigh. The outflow trajectories became more westerly as they approached the northern edge of the anvil resulting from Coriolis accelerations acting over several hours.

Strong simulated upper-level outflow occurred over a broad zone of several hundred kilometers north and east of the MCS convection as found by Fritsch and Maddox (1981). However, the simulated TKE was most concentrated over an ~150-km ring-shaped region extending southward from the north anvil edge. Here, upper-level outflow was both stronger and more vertically concentrated than it was closer to the convection, which resulted in stronger vertical shears at aircraft flight levels beneath the outflow.

The model used in this study lacks sufficient resolution to explicitly simulate turbulence and the observations are insufficient to determine the precise mechanisms of the turbulence production. Nevertheless, the simulations do suggest that strong vertical shear in the MCS upper-level outflow can promote turbulence over a mesoscale region well downstream of the outflow source by supporting dynamic (e.g., Kelvin–Helmholtz) instabilities. Furthermore, it was found that this enhanced vertical shear also acted to modify the static stability by differentially advecting horizontal (equivalent) potential temperature gradients influenced by adiabatic cooling associated with the mesoscale updraft in the upstream (southward) portion of the anvil. This resulted in a steepening of the isentropes between the mesoscale updraft and the northern anvil cloud edge and the development of localized regions of moist static instability.

The largest-amplitude fluctuations in simulated moist static stability and related variations in Ri and TKE had periods of several hours and appear to be related to low-frequency intensification and decay of mesoscale components of the convective system, which in turn influenced properties of the upper-level outflow. Thus, our proposed turbulence mechanism is distinct from those discussed in previous model-based convectively induced turbulence studies in which the CIT was either directly associated with high-frequency gravity wave breaking (e.g., Lane et al. 2003; Lane and Sharman 2008) or with more localized reductions of Ri resulting from horizontally propagating gravity waves (e.g., Fovell et al. 2007).
Comparison of the control simulation with the “adiabatic” simulation confirmed that strong vertical shear, and reduced Ri at flight levels along the northern anvil edge in the control simulation resulted almost entirely from the MCS-induced upper-level outflow circulation. This contrasts with the south and southwest edge of the MCS anvil where outflow easterly shear was significantly compensated for by environmental westerly shear, which resulted in larger Ri and less widespread simulated TKE than at the northern anvil edge.

The pattern of deep convection and its relationship to the environmental winds influences the MCS outflow, its associated vertical shear, and thus the potential for CIT. Figure 20 summarizes relationships among these features, which themselves are modified somewhat to better represent established climatological archetypes. For example, Schumacher and Johnson (2005) discuss similar heavy-rain-producing training line-adjoining stratiform MCSs that comprise west–east-oriented convective lines with an asymmetric divergent upper-level outflow located primarily to the north of the convection. They show that such precipitation systems occur in similar veering environmental winds characterized by 1) a southwesterly low-level jet of high-$\theta_e$ air that feeds the convection on the south part of the MCS cloud shield and 2) significant upper-level environmental westerlies that may augment the upper-level outflow downstream.

In the current case, the MCS matured along and slightly north of the environmental upper-level jet, which contrasts with a 4-yr warm-season RUC-based climatology in which heavy rainfall systems were located several hundred kilometers south of the 200-hPa environmental jet (Fig. 1). However, in both the current case and the climatology, the strongest upper-level outflow is in the direction of the environmental westerlies, with the maximum winds located several hundred kilometers north of the heaviest rainfall and occurring several hours later. These spatiotemporal relationships between the heavy rain and upper-level outflow have implications for forecasting most likely locations of widespread flight-level turbulence, given the strong vertical shear in both the MCS outflow and the midlatitude storm environment at or immediately below the tropopause, where MCSs typically detrain (e.g., Maddox 1983; Cotton et al. 1989).

An encouraging aspect of the current case study was the ability of the RUC operational model analysis to capture the signature of the MCS upper-level outflow and the associated low background Ri values of $\sim 1$ at aircraft flight levels, which reflects the mesoscale nature of the flow destabilization. Such lowering of Ri is likely to be common in the upper-level outflows of MCSs and therefore may be a useful concept for developing empirically based algorithms of CIT likelihood.

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![Fig. 20. Schematic diagram illustrating spatial relationships among the environmental flow, strongest deep convection (thunderstorm symbols), MCS anvil cloud (thin oval-shaped curve), MCS divergent upper-level outflow streamlines (thick curves with arrowheads), and the location of strongest upper-level outflow winds (gray shading).](image-url)