Mean Flow and Turbulence Characteristics of a Nocturnal Downburst Recorded on a 213-m Tall Meteorological Tower

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ABSTRACT: This study presents rare measurements and analysis of a nocturnal thunderstorm downburst on the 213-m tall Cabauw tower in the Netherlands. The event occurred between 0200 and 0300 UTC 12 March 2008 and was measured using four ultrasonic 10-Hz anemometers positioned at 3, 60, 100, and 180 m above ground level. One-second gusts in the outflow exceeded 30 m s^{-1} at 60 m and above. This wind event was accompanied by an abrupt change of wind direction from southwest to west. While the shift in wind direction corresponded with the change of upwind surface roughness, the time series of turbulence intensity and other turbulence characteristics were not affected. The statistical properties of this event were compared against the largest European database of thunderstorm winds measured in the Mediterranean. The study also demonstrated that primary and secondary vortex structures—secondary vortex being rarely observed in actual downbursts—developed at the forward edge of the cold outflow. The estimated diameter of the downdraft was 1200 m at 70 m above ground level. The measured velocity profiles and friction velocity were compared against theoretical predictions of the Monin–Obukhov similarity theory (MOST). MOST without stratification adjustment overestimated measured friction velocity twofold. Alternative values for surface roughness during the outflow were derived based on the measured friction velocity and MOST-based fit of measured velocity profiles. Ceilometer and radar measurements were supplementary data in this analysis.

SIGNIFICANCE STATEMENT: We analyzed the occurrence of a strong nocturnal downburst measured on a 213-m tall meteorological tower in the Netherlands (0200–0300 UTC 12 March 2008). The near-surface (3-m) wind gusts in the outflow exceeded 20 m s^{-1} . We showed the existence of the primary and secondary circulation vortices at the leading edge of the cold outflow. The secondary vortex that is smaller and formed ahead of the primary vortex is rarely observed in actual events. The classical approach used in numerical weather prediction models to estimate the influence of shear stress on the mean flow overestimated the friction velocity twofold. Future studies should focus on more cases of nocturnal thunderstorm winds.

KEYWORDS: Europe; Gust fronts; Turbulence; Downbursts; Thunderstorms; Wind; Wind gusts; Measurements

1. Introduction

Thunderstorm winds have been one of the main research topics in atmospheric sciences and wind engineering over the last several decades. Downbursts are a class of thunderstorm winds defined as strong downdrafts of cold air that originate in the cloud and spread out radially after impinging on the surface (Fujita and Byers 1977). The main contributors to downdraft development are the evaporation of hydrometeors inside and underneath the cloud, as well as the melting of ice and drag exerted by the falling hydrometeors. A vertical nonhydrostatic pressure gradient term can also be substantial in many non-single-cell-type thunderstorms. The increase of perturbation pressure with height provides a downward force on the air parcel.

Due to their unpredictability, often intense magnitudes, highly three-dimensional flow field, and non-Gaussian velocity

distributions (Hangan et al. 2019), downbursts are dangerous for aircraft during their landing and takeoff maneuvers (Fujita and Caracena 1977; Fujita and Byers 1977; Pryor 2015), as well as low-rise structures (Abd-Elaal et al. 2018; Lombardo et al. 2018). However, high-frequency measurements ($>1 \text{ Hz}$) of downbursts are still infrequent in comparison to their measurements at weather stations using cup anemometers. Besides, various financial and organizational challenges associated with full-scale measurements usually prevent many field campaigns. For instance, thunderstorm field campaigns often require rapidly deployable adaptive observing networks which are costly and challenging to operate. Particularly rare are the high-frequency measurements of downbursts on tall meteorological towers whose heights exceed 100 m above ground level (AGL; Table 1). While the list might not be exhaustive, Table 1 shows that most field campaigns prior to 2000 measured thunderstorm winds using low-frequency anemometers. Those data are therefore of limited use in the investigation of turbulence and transient

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TABLE 1. Summary of thunderstorm wind measurements from tall (≥ 100 m AGL) meteorological towers.

Study	Geographical region	Tower height (m) (number of measuring heights)	Anemometer sampling frequency (Hz)	Study's research focus
Goldman and Sloss (1969)	United States	150 (6)	1	Structure of the leading edge of thunderstorm outflow
Sinclair et al. (1973)	United States	150 (6)	10	Variation of the low-level winds during the passage of a thunderstorm gust front
Charba (1974)	United States	444 (7)	0.1	Detail analysis on the relationship between gravity current models and squall-line gust front
Goff (1975)	United States	444 (7)	0.1	Detail analysis of 20 thunderstorm outflows captured by anemometer and Doppler radar data
Goff (1976)	United States	444 (7)	0.1	As in Goff (1975)
Hall et al. (1976)	United States	150 (6)	1	Wind shear analysis of thunderstorm outflow using anemometers and acoustic sounder
Shapiro (1984)	United States	300 (8)	10	Structure and physical processes of an intense cold front
Beljaars (1987)	The Netherlands	200 (6)	2	The influence of sampling and filtering times on measured wind gusts
Sherman (1987)	Australia	200 (4)	1	Outflow characteristics of a downburst using in situ and Doppler radar measurements
Johnson et al. (1987)	United States	444 (7)	0.1	Initiation of rotation at mid- and low levels, and various thermodynamic analyses of a tornadic storm
Intrieri et al. (1989)	United States	300 (3)	10	Analysis of colliding thunderstorm outflows using anemometer and remote sensing techniques
Dowell and Bluestein (1997)	United States	444 (7)	0.67	Dynamics of low-level inflow, updraft, mesocyclone, and rear precipitation core of a supercell that produced an F2 tornado
Choi (2004)	Singapore	152 (5)	50	Variation of wind velocity with height during more than 50 thunderstorms; influence of surface roughness
Lombardo et al. (2014)	United States	200 (10)	30	Investigation of thunderstorm velocity records in the wind engineering framework
Gunter et al. (2015)	United States	200 (10)	50	Validation of wind profiles derived from dual-Doppler radars (thunderstorm and nonthunderstorm events analyzed)
Gunter and Schroeder (2015)	United States	200 (8)	50	In situ and Doppler radar measurements of several thunderstorm wind events in terms of their structure, evolution, and variability
Zhang et al. (2019)	China	325 (15) (9)	(0.05)(10)	Characteristics of thunderstorm outflows in Beijing, China

characteristics of downbursts. Also, some of the studies that used the data from tall meteorological towers were focused on an entirely different subject, such as the thunderstorm dynamics at meso- α (200–2000 km) and meso- β (20–200 km) scales or the process of tornadogenesis. Therefore, little consideration was given to downburst outflows.

This scarcity of downburst measurements from tall meteorological towers is one of the reasons for the slower pace of theoretical development of thunderstorm boundary layer

dynamics in comparison to reasonably well-established theories that hold for the large-scale atmospheric boundary layer (ABL) winds ([Stull 1988](#)). From the micrometeorological point of view, the main difficulty in the investigation of downbursts is their nonstationarity and significant variability of the flow field in space ([Orwig and Schroeder 2007](#); [Gunter et al. 2017](#)). The unsteadiness and flow inhomogeneity limit the application of the Monin–Obukhov similarity theory (MOST) to the framework of downburst winds ([Markowski et al. 2019](#)). MOST is a

generalization of the law of the wall that governs the surface layer fluxes and velocity profiles in a homogeneous fluid to the case of a stratified fluid, such as Earth's atmosphere. While this theory is the backbone of surface layer parameterization in most numerical models, the applicability of MOST to thunderstorm winds and downbursts is incomplete, to say the least. The present study attempts to test the applicability of the MOST framework to transient winds by fitting the observed velocity profiles with MOST predictions using a range of Obukhov lengths. We compared measured friction velocities and the MOST-derived values at different heights in the outflow.

The downburst analyzed in this study occurred during night when a nocturnal boundary layer (NBL) was present. The stable NBLs are more delicate to analyze than the daytime ABLs due to the common absence of stationarity and homogeneity of turbulence (Caughey et al. 1979; Nieuwstadt 1984). As John C. Wyngaard pointed out, "The stable boundary layer is as different from the convective boundary layer as night is from day" (Wyngaard 2010, p. 267). In NBL, turbulence is generated only through shear production and, since the buoyancy predominantly affects only the large eddies, the lack of convection in nighttime terminates the energy cascade and diminishes the turbulence. Therefore, the turbulence in NBL is very sensitive to even small changes in atmospheric stability, which hinders the establishment of the boundary layer equilibrium and, hence, complicates the formulation of a unified theory of NBL turbulence (Zilitinkevich and Baklanov 2002).

The goal of this paper is to investigate the mean and turbulent flow characteristics of a nocturnal downburst measured on the tall meteorological tower located in the Netherlands—Cabauw tower. While this downburst was nocturnal, thunderstorm winds are more frequent in the afternoon. Zhang et al. (2019) found that only around 6% out of 70 analyzed thunderstorm winds in Beijing, China, occurred between 0000 and 0600 local time. While the percentage was higher (20%) in Burlando et al. (2018) because most of their events were associated with thunderstorms that developed above the sea and moved toward the coast, the period 0000–0600 local time was still the least likely part of the day for thunderstorms development. The results presented in this study, therefore, provide unique insights into the spatiotemporal evolution of nocturnal downbursts and their velocity profiles in this part of Europe.

2. Data and methodology

a. Wind measurements

The wind data come from the 213-m tall Cabauw tower located in the Netherlands. This meteorological tower is situated in a polder 0.7 m below mean sea level in the west part of the Netherlands (latitude: 51°58'12.00"N, longitude: 4°55'34.48"E) (Bosveld et al. 2020). The site is flat in all directions within ~20 km from the tower. The land use within ~400 m around the tower is open pasture (Petrović et al. 2018; van den Berg 2008; Verkaik and Holtslag 2007).

The wind data used in this study were measured at 3, 60, 100, and 180 m AGL during the 1-yr period from 1 July 2007 to

FIG. 1. The Cabauw site near the town of Lopik in the Netherlands: the 213-m tall meteorological tower (circle), the 3-m mast (square), and the ceilometer (triangle). Source of background map: Google Earth images.

30 June 2008. The 3-m measurements are not from the main tower, but from an auxiliary mast that was approximately 200 m north from Cabauw tower (Fig. 1). The wind measurements were conducted using the triaxis Gill R3 sonic anemometers (Solent 20-cm path sonic anemometer at 100 m) with a sampling frequency of 10 Hz and a measuring volume of 0.15 m. The precision of wind speed and direction measurements is 0.01 m s^{-1} and 1° , and their range is $0\text{--}45 \text{ m s}^{-1}$ and $0^\circ\text{--}359^\circ$, respectively. The accuracy of reliable wind measurements that pass the quality check is $<1\%$ relative root-mean-square error. Some electrical spikes were detected during the investigated wind event that occurred in rainy conditions. The data are highly experimental and were provided to the author "as is" with only standard calibrations applied to raw measurements (Bosveld 2016; see appendixes A and B). For wind directions between 280° and 340° , the sensors at 60, 100, and 180 m were in the wake of the tower and those data are less reliable than for other wind directions. The length of the booms that hold the instruments is 9.4 m and the booms are pointed toward the southeast. The anemometers were located at the end of the booms. With this boom length, the interference by the tower is $\sim 1\%$ at the upwind side (Wyngaard 1981). Also, the sensors were installed at 0.8 m above the booms to minimize data contamination due to the booms themselves. Fuertes et al. (2014), de Roode et al. (2010), and Baltink et al. (2009) used the same 10-Hz data in their analyses of turbulent fluxes, wind dynamics, and validation of lidar measurements.

b. Auxiliary observations

The Vaisala CT75 ceilometer (Fig. 1) was used to estimate the cloud-base height and backscatter profile. CT75 is a low-power lidar that transmits short light pulses with a wavelength of 905 nm and collects the attenuated backscatter coefficient (β) from clouds and aerosols as a function of height. This ceilometer can detect up to three cloud bases every 30 s with the vertical range resolution of backscatter and cloud base of

30 and 15 m, respectively. The cloud bases are retrieved using Klett's (1981) inversion of the β profile assuming a fixed extinction-to-backscatter ratio while taking into account the slope, absolute value, and historic observations at that height.¹ The ceilometer was continuously scanning vertically with an elevation of 88° above the horizon. The maximum detection height was 11.25 km above the surface.

In addition to ceilometer measurements, radar reflectivity from a C-band (5.6-GHz) pulsed radar in De Bilt, the Netherlands, was used to analyze the spatiotemporal characteristics of the squall line that produced the downburst. These data were also used to determine storm motion. The maximum range of the De Bilt radar is 320 km with a gate resolution of 1 km. The temporal resolution and beamwidth are 5 min and 1°, respectively. The scanning elevations are in the range of 0.3°–25° and ground clutter correction was operationally applied. The radar in De Bilt is situated approximately 22.8 km in the north-northeast direction from the Cabauw site. Last, the standard 10-min meteorological measurements of temperature, pressure, and precipitation that were recorded at the tower are discussed in conjunction with high-frequency wind measurements.

c. Downburst detection and signal despiking

A four-step downburst detection algorithm proposed by Fujita (1985) was used to detect downbursts in the year-long velocity measurements. First, left-sided 1-min moving means $\bar{u}_{1\min}$ were computed from the instantaneous velocity (u) as

$$\bar{u}_{1\min}(t) = \frac{1}{k+1} \sum_{j=0}^k u(t-j), \quad (1)$$

where t is the time, $k = 60f_s$ is a 1-min moving window, and f_s is the anemometer sampling frequency. If the following conditions are satisfied simultaneously,

$$\bar{u}_{1\min} > 10 \text{ m s}^{-1}, \quad (2a)$$

$$\bar{u}_{1\min} > \begin{cases} \bar{u}_+ + 5 \text{ m s}^{-1} \\ \bar{u}_- + 5 \text{ m s}^{-1} \end{cases}, \quad (2b)$$

$$\bar{u}_{1\min} > \begin{cases} 1.25\bar{u}_+ \\ 1.25\bar{u}_- \end{cases}, \quad (2c)$$

$$\bar{u}_+ \leq 1.5\bar{u}_-, \quad (2d)$$

the event is classified as a downburst. Here, $\bar{u}_-$ and $\bar{u}_+$ are the pre- and postpeak mean velocities calculated from seven 1-min segments before and after a $\bar{u}_{1\min} > 10 \text{ m s}^{-1}$ peak, respectively. However, the first minute before and after the peak was excluded from the calculations of $\bar{u}_-$ and $\bar{u}_+$. Equation (2b) ensures that the peak velocity is at least 5 m s^{-1} higher than the background winds before and after the event. The condition

(2c) prevents the identification of small bell-shaped profiles that might be due to a random turbulent fluctuation in the wind. Equation (2d) dismisses wind events associated with strong and persistent winds after the frontal passage (Wakimoto 1985). Fujita (1985) extracted 579 microbursts from the Northern Illinois Meteorological Research on Microburst (NIMROD) and Joint Airport Weather Studies (JAWS) field campaigns combined using this procedure.

It was necessary to subject all wind measurements to quality control and removal of electronic spikes. The spikes in sonic anemometer readings usually occur during precipitation when water collects on the transducers (Vickers and Mahrt 1997). The first quality check was based on the absolute value of velocity difference between two consecutive readings. If the difference was larger than a scaling factor α or if the difference was zero, the point was flagged as a spike and removed. The proper value of α was empirically determined to be 4. The inclusion of zero difference in the method flags the spikes that sometimes plateau at the fixed value over a short period. In addition to this quality check, we also implemented a Savitzky–Golay filter for despiking of noisy data (Savitzky and Golay 1964). This method is a least squares regression of a set of consecutive data points to a higher-order polynomial. The central point of the fitted polynomial curve, U_i ($i = 0, 1, 2, \dots$), is a new value of smoothed velocity:

$$U_i = \sum_{n=-n_L}^{n_R} c_n u_{i+n}, \quad (3)$$

where $u_i \equiv u(t_i)$ is a set of velocity data equally spaced in time $t_i = t_0 + i$, and n_L and n_R are the number of points left and right of i , respectively. The convolution coefficients c_n were found as the least squares fit of velocity data (Press et al. 2007). We employed a quadratic polynomial fit with the moving window of $5f_s - 1 = 49$ points.

In addition to instantaneous velocity, the Savitzky–Golay filter was also applied to the reduced turbulent fluctuations (see section 2c for the definition of reduced turbulence). Any point in either of these two records was flagged as a spike if

$$|U_i - u_i| > \lambda_u, \quad (4a)$$

$$|\tilde{U}'_i - \tilde{u}'_i| > \lambda_{u'}, \quad (4b)$$

where $\lambda_u = 1.9$ and $\lambda_{u'} = 4.0$ are the threshold values for instantaneous velocity (u) and the reduced turbulent fluctuations (u'), respectively. All model parameters were determined based on visual inspection of data (Vickers and Mahrt 1997) and spectral analysis of the reduced turbulence fluctuations (see appendix A). Any measurement that was flagged as a spike was removed from the data. We do not recommend the interpolation of data gaps due to the highly transient nature of downbursts and the high frequency of raw measurements. While different interpolation techniques are justifiable in the case of, say, average 10-min winds in fair weather conditions, it is not clear if they are adequate—likely not—in the case of transient and non-Gaussian downbursts. The results of the proposed despiking method are presented in appendix A.

¹ The exact algorithm is regarded as commercially sensitive by Vaisala and has not been released.

d. Theoretical background

The vertical profile of mean horizontal velocity $\bar{u}_H$ in the framework of the MOST is expressed as (Businger et al. 1971)

$$\bar{u}_H(z) = \frac{u_*}{\kappa} \left[\ln\left(\frac{z}{z_0}\right) - \Psi_1\left(\frac{z}{L}\right) \right], \quad (5)$$

where u_* is the friction velocity, z_0 is the roughness length, κ is the von Kármán constant (0.4), and z is the height AGL. Also, $\bar{u}_H = (\bar{u}^2 + \bar{v}^2)^{1/2}$, where $\bar{u}$ and $\bar{v}$ are velocity components in the horizontal plane. The stratification correction term $\Psi_1(z/L)$ is adopted from the field experiments reported in Businger et al. (1971):

$$\Psi_1\left(\frac{z}{L}\right) = \begin{cases} 2 \ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - \arctan(x) + \frac{\pi}{2}, & L \leq 0 \\ -4.7 \frac{z}{L}, & L > 0, \end{cases} \quad (6)$$

where

$$x = \left(1 - 15 \frac{z}{L}\right)^{1/4}. \quad (7)$$

The governing geometric parameters of the flow in the MOST framework are z and the Obukhov length L that is defined as

$$L = -\frac{u_*^3 \bar{T}}{\kappa g w' \bar{T}'}, \quad (8)$$

where $\bar{T}$ is the average surface layer temperature, $w' \bar{T}'$ is the heat flux (w is the vertical velocity), and $g = 9.81 \text{ m s}^{-2}$ is the gravitational acceleration. Hereafter, an overbar denotes a time average and a prime is a deviation from the average.

The choice of proper averaging time is a challenging question in the investigation of transient wind events such as downbursts. The averaging time of 15 min was chosen in the original work of Businger et al. (1971), but their study did not consider thunderstorm winds. More recently, Markowski et al. (2019) discussed the appropriate averaging time in the case of thunderstorms and proposed a 30-min averaging period before the arrival of cool outflow and 15 min within the outflow. The authors further recognized the importance of this issue and stated, “it is unclear what averaging period is ideal.” This question was also addressed by Beljaars (1987) in his study on the proper filtering time of gust measurements, as well as by Holmes et al. (2008), Lombardo et al. (2014), and Solari et al. (2015), among others. The present study uses 20-min averaging intervals of slowly varying mean velocity records (i.e., instantaneous data were first smoothened using a 30-s moving-mean operator). The details are discussed below and in section 3c.

The instantaneous velocity of a transient wind event can be decomposed via (e.g., Holmes et al. 2008; Choi and Hidayat 2002; Chen and Letchford 2004; Bendat and Piersol 2010; Solari et al. 2015; Romanic et al. 2020b)

$$s(z, t) = \bar{s}(z, t) + s'(z, t), \quad (9)$$

where $s \equiv (u, v, w)$. This expression represents the generalization of classical Reynolds decomposition of stationary winds to transient winds. The slowly varying mean $\bar{s}(z, t)$ is associated with the large-scale features of downburst outflow and obtained by applying a moving average filter in the form of Eq. (1) to instantaneous data. In downbursts, $s = s(\mathbf{r}, t)$, where $\mathbf{r}$ is the position vector in three dimensions, but we omit this notation because the present measurements are restricted to the vertical tower and the three-dimensional variability of the entire outflow cannot be assessed.

Since fluctuations $s'(z, t)$ are also transient process in downbursts, this term is further expressed as (Solari et al. 2015)

$$s'(z, t) = \sigma_s(t, z) \tilde{u}_s(t, z), \quad (10)$$

where σ_s is the slowly varying standard deviation of s , and $\tilde{u}_s$ is the reduced turbulence fluctuations, which are a stationary and Gaussian process. By combining Eqs. (9) and (10), we arrive with the final expression for the decomposed velocity:

$$s(z, t) = \bar{s}(z, t) [1 + \bar{I}(z, t) \tilde{u}_s(t, z)], \quad (11)$$

where $\bar{I}(z, t) = \sigma_s(t, z) / \bar{s}(z, t)$ is the slowly varying turbulence intensity.

The velocity decomposition was performed using a 30-s averaging time (t_{ave}) (Burlando et al. 2017). Solari et al. (2015) demonstrated that this averaging period accurately disjoints the spectral content of low-frequency motions, $\bar{s}(z, t)$, from the high-frequency fluctuations, $\tilde{u}_s(t, z)$. It will be demonstrated in section 3a that $t_{\text{ave}} = 30 \text{ s}$ also results in $\tilde{u}_s(t, z)$ being a Gaussian and stationary process that follows the $-5/3$ slope of turbulence spectra in the inertial subrange. Junayed et al. (2019) and Romanic et al. (2020b) showed that the velocity records of physically produced downburst-like outflows in a wind simulator are similar to the actual thunderstorm events that are smoothed with the 30-s averaging filter (Holmes et al. 2008; Lombardo et al. 2014).

Eddy covariance observations at four heights on the tower were used to calculate u_* :

$$u_* = \left[(\overline{w'u'})^2 + (\overline{w'v'})^2 \right]^{1/4}, \quad (12)$$

where $\overline{w'u'}$ and $\overline{w'v'}$ are the vertical momentum fluxes in the streamwise and crosswise directions, respectively. However, to obtain the proper streamwise and crosswise directions in the highly three-dimensional downburst, it is necessary to move from the anemometer frame of reference to the natural coordinates that are referenced to the mean velocity vector. This rotation of coordinates is particularly important in the case of downbursts that are almost always associated with sharp changes in wind direction (Romanic et al. 2020a). Appendix B formally describes the rotation of wind components and fluxes from the anemometer to natural coordinates.

3. Results

a. Velocity records and signal analysis

Downburst measurements in Fig. 2 are transient velocity records with the gusts exceeding 30 m s^{-1} at $\geq 60 \text{ m}$ and abrupt

FIG. 2. Instantaneous horizontal velocity (u_H , black lines and the left ordinate) and the slowly varying mean of wind direction (α , gray dots and the right ordinate) during the downburst passage over the Cabauw site on 12 Mar 2008. The anemometers were positioned at four different heights AGL (z_A). The arrows above each plot indicate the winter value of surface roughness (z_0) for different wind directional sectors (Table 2).

change of wind direction at all levels. Using the European Severe Weather Database terminology, this event can be classified as a severe wind due to the gusts exceeding 25 m s^{-1} . The downburst was accompanied by a sudden drop of temperature ($\sim 4^\circ\text{C}$) and a pressure jump of $\sim 0.6 \text{ hPa}$ (Fig. 3). The moderate precipitation during the event exceeded 0.7 mm over 10 min . While the surface pressure and temperature were only available as 10-min means, the pressure drop that preceded the temperature decline and precipitation is a known precursor of the approaching thunderstorm. Strong winds at the upper three levels exceeded 20 m s^{-1} and lasted for $\sim 12 \text{ min}$, between 0227 and 0239 UTC. The velocities at 3 m were approximately 10 m s^{-1} lower than at the higher levels.

The change of wind direction for $\sim 35^\circ\text{--}40^\circ$ at all heights was associated with the passage of a squall line (section 3b) and the directional shift preceded the velocity ramp up (Fig. 2). The wind directional switch between two different regimes that persisted over a longer time period than the downburst duration (Fig. 4) shows that the event was associated with the frontal passage rather than an isolated thunderstorm. Further discussion of weather at larger scales is provided in section 3b. Here, we continue with the analysis

of wind records and notice that a visually similar wind event occurred around 0450 UTC on the same day (Fig. 4), but Fujita's algorithm did not identify it as a downburst.

FIG. 3. (top) The 10-min values of the surface air temperature (T , dots and left ordinate) and mean sea level pressure (p , stars and right ordinate) and (bottom) the accumulated precipitation (P). The nonzero values of precipitation rates are provided for each bar.

FIG. 4. Six hours of instantaneous horizontal velocity (u_H , black lines and the left ordinate) and wind direction (α , gray dots and the right ordinate) at 100 m on 12 Mar 2008. The shaded region between 0200 and 0300 UTC is the 1-h interval centered on the downburst analyzed in this study (Fig. 2).

Figure 4 suggests that transient winds similar to the one analyzed here are frequently observed at the tower.

Figure 2 further shows the relationship between the changing wind direction and climatological z_0 values for different directional sectors. Since the downburst occurred in March, the winter values of z_0 around the site (Beljaars and Bosveld 1997) were deemed more appropriate for this analysis than the summer figures (Table 2). The timing of peak velocity in the outflow and the change of wind direction corresponded with the change of z_0 of these two wind directional sectors. Although the change of z_0 from 0.04 to 0.03 m is a minor difference even for ABL winds, the turbulence characteristics in the outflow were not largely affected (Fig. 5). Turbulent fluctuations (u'_H) and turbulence intensity (I_{uH}) either slightly decreased or were not affected when the wind direction shifted to the lower z_0 . The weak dependency of thunderstorm winds on the changes of z_0 was also noticed by Burlando et al. (2017) and Romanic et al. (2020a). Because downbursts are highly transient, the developing boundary layer seems not to reach an equilibrium with the underlying surface, as it does in the case of large-scale and steady ABL winds.

The velocity peak during the downburst is observed in the slowly varying mean velocity records at all heights (Figs. 5a–d). This transient feature of the mean flow confirms that the phenomena could not be a random fluctuation, but a larger-scale wind system (Fig. 4) that was also characterized by transient and non-Gaussian turbulent fluctuations (Figs. 5e–h). The assumption of constant I_{uH} that is sometimes invoked in the literature (Kwon and Kareem 2009) is questionable in this case, particularly at $z_A = 180$ m (Fig. 5i). We observe a decrease of I_{uH} at all heights after the downburst passed over the tower. A similar trend of I_{uH} is often reported for thunderstorm outflows even during daytime events (Holmes et al. 2008; Burlando et al. 2017; Romanic et al. 2020a).

The power spectral density ($S_{u'_H}$) and other characteristics of reduced turbulent fluctuations ($\tilde{u}'_H$) are shown in Fig. 6. The spectra were calculated using Welch's (1967) method of modified periodograms with the spectral window of 2^{11} readings and a 50% overlap between the adjacent windows. A Hamming window was used to modify periodograms and reduce spectral leaking at the boundaries. At all heights, the records of $\tilde{u}'_H$ resemble a stationary and Gaussian process (Figs. 6a–h, Table 3), and their spectra follow the $-5/3$ slope

in the inertial subrange. The slight flattening of $S_{u'_H}$ in the range of highest frequencies is caused by the noise in ultrasonic anemometers (appendix A). In the low-frequency domain, the spectral peak at 3 m is shifted for approximately 0.2 Hz to higher frequencies in comparison to the upper three levels. This shift of peak frequency suggests larger integral length scales (l'_{uH}) at higher elevations knowing that both l'_{uH} and the peak of $S_{u'_H}$ measure the size of turbulence eddies (Peña et al. 2010).

Indeed, the values of l'_{uH} , which are not obtained from spectra but calculated from normalized autocorrelation functions of $\tilde{u}'_H$ (Flay and Stevenson 1988), show that the integral length scales at higher elevations are about 2 times larger than at 3 m (Table 3). In this analysis, the autocorrelation functions were normalized with the variance of $\tilde{u}'_H$ and the integration was performed from 0 to the time lag at which the autocorrelation reduces to 0.05. Then, using Taylor's hypothesis of frozen turbulence, we computed $l'_{uH} = \tau_{\tilde{u}'_H} \cdot \hat{u}_H$, where $\tau_{\tilde{u}'_H}$ is the time scale of turbulence obtained from the autocorrelation function and $\hat{u}_H$ is the maximum value of the slowly varying velocity. Only the 10-min interval around the peak velocity was used in this analysis. The values of l'_{uH} during the downburst passage (Table 3) fall between the values of l calculated for the ABL winds under the neutral and near-unstable atmosphere in Peña et al. (2010). However, the mean velocities in the 10-min period during the downburst passage were above 13 m s^{-1} (except at 3 m) and this interval is outside the considered range of mean velocities in Peña et al. (2010). Furthermore, they computed l using a different methodology.

We further compare the investigated downburst against a “mean” downburst derived from the largest database of thunderstorm wind measurements in Europe (Zhang et al. 2018; Table 3). Their field campaign used 28 ultrasonic anemometers

TABLE 2. Roughness length (z_0) around the Cabauw site (Beljaars and Bosveld 1997). This table only provides the values of z_0 that are relevant for the range of wind directions in Fig. 2.

	Wind direction sector		
	220°–240°	240°–260°	260°–280°
Winter z_0 (m)	0.02	0.04	0.03
Summer z_0 (m)	0.04	0.07	0.06

FIG. 5. Rows show the decomposed horizontal velocity from all anemometers. Columns show (a)–(d) slowly varying mean velocity ($\bar{u}_H$), (e)–(h) turbulent fluctuations (u'_H), and (i)–(l) slowly varying turbulence intensity (I_{uH}).

installed along the coast of the Ligurian and the northern Tyrrhenian Seas in the Mediterranean (over 270 transient wind records). Knowing that their measurements were mostly restricted to the heights between 10 and 30 m, it is perhaps not surprising that their gust factor (G) nicely corroborates with the value obtained in this study at 3 m. The gust factor is defined as $G = \hat{u}_{1s,H} / \bar{u}_H$, where $\hat{u}_{1s,H}$ is the 1-s peak velocity and $\bar{u}_H$ is the maximum value of the slowly varying wind speed. Most thunderstorms in Zhang et al. (2018) approached the anemometers

from upstream fetch over water (Burlando et al. 2017, 2018), and since G in principle depends on z_0 , one would expect that the values of G in their study would differ from those found in our paper. However, such discrepancy is not observed in Table 3. Lombardo et al. (2014) reported similar values of G for an entirely different site in the United States and different types of thunderstorm winds. The values of G during thunderstorm winds are not noticeably affected by z_0 , type of thunderstorm winds and the time of the day when the winds occurred.

FIG. 6. (a)–(d) Reduced turbulent fluctuations, (e)–(h) their histograms, and (i)–(l) their power spectral densities at all heights.

The decreasing trend of G with the height is interrupted at 180 m, and G returns to the value observed at 3 m (Table 3). We speculate that this trend of G could be due to the nose-shape profile of mean velocity in the downburst (section 3c) combined with monotonically increasing $\hat{u}_{1s,H}$ with the height. The positive feedback between G and $\hat{u}_H$ was already suggested in Zhang et al. (2018), and our study supports the hypothesis by providing the data from a tall tower and a different site. On the other hand, the wind direction at 180 m appears to veer more to 280° during the peak of the outflow

(Fig. 2). That direction is close to the indicated azimuths (280° – 340°) that were influenced by the tower, which also might increase turbulence, and thereby increase G and reduce $l_{u'_H}$.

b. Downburst outflow structure and evolution

The vertical structure of downburst outflows and the nose-shape velocity profile have been reported in many field observations (Byers and Braham 1949; Fujita 1985; Goldman and Sloss 1969; Gunter and Schroeder 2015), and their structure

TABLE 3. Flow parameters of the downburst investigated in this study and their comparison against the largest European downburst database (Zhang et al. 2018). The parameters are evaluated for a 10-min period centered on the time instance of maximum slowly varying mean velocity ($\bar{u}_H$). See text for the meaning of different parameters.

Parameter	Anemometer height (m)				Zhang et al. (2018)
	3	60	100	180	
$\hat{u}_{1s,H}$ (m s ⁻¹)	21.40	30.45	31.33	33.88	33.65
$\hat{u}_H$ (m s ⁻¹)	17.88	28.12	29.60	28.68	27.68
G	1.20	1.08	1.06	1.18	1.21
$\bar{I}_{uH}$ (%)	15.89	7.60	7.25	6.17	12.0
$\mu_{\hat{u}_H}$	0.01	-0.01	-0.01	0.00	-0.01
$\sigma_{\hat{u}_H}$	0.99	0.89	0.92	0.93	1.01
$k_{\hat{u}_H}$	2.49	3.22	2.83	3.35	2.84
$\gamma_{\hat{u}_H}$	0.01	-0.13	0.06	-0.16	-0.10
$l_{\hat{u}_H}$ (m)	22.33	41.19	56.20	38.84	28.53

was studied in laboratory tank experiments with density currents (Keulegan 1958; Charba 1974; Lundgren et al. 1992) and impinging jets (Romanic and Hangan 2020). While a secondary vortex (SV) that precedes the primary vortex (PV) is readily observed in the physical and numerical simulations of downbursts (Lundgren et al. 1992; Yao and Lundgren 1996; Kim and Hangan 2007; Mason et al. 2010; Vermeire et al. 2011; Orf et al. 2014) (Fig. 7), this dynamical feature of the outflow has only been reported in a few full-scale measurements (Sherman 1987; Chai et al. 2004).

The existence of SV and PV is observed after careful inspection of the time series of vertical velocity (w) in Fig. 8 and their relationship to the time histories of $\bar{u}_H$ in Fig. 5a–d. The PV that developed along the leading edge of the outflow is the main signature of all downburst velocity records. In terms of the slowly varying vertical velocity ($\bar{w}$), the PV is characterized by the steady increase of $\bar{w} > 0$ that preceded the ramp up of $\bar{u}_H$, as schematized in Fig. 7. The peak value of $\bar{w}$ (i.e., $\hat{\bar{w}}$) occurred ~ 100 s before $\hat{\bar{u}}_H$. However, the strong positive $\bar{w}$ is pronounced only at higher elevations in the PV (Fig. 8a) due to the size of PV that is typically several hundreds of meters to over 2 km high (Mueller and Carbone 1987). While the increase of $\bar{u}_H$ is observed at all heights (Figs. 5a–d), the PV produced a strong $\bar{w} > 0$ only at higher levels. Also, $\hat{\bar{w}}$ is observed at a higher elevation than the height of the maximum $\hat{\bar{u}}_H$, which is usually between 80 and 150 m (Fig. 7 and Table 3). This observation agrees well with lidar measurements of Canepa et al. (2020), where the authors showed the strong upward motions below ~ 100 m in the PV are rare. The overturning flow at the rear side of the PV (Fig. 7) is expected to produce a negative $\bar{w}$, which is indeed observed in Fig. 8a around 0231 UTC.

Another prominent feature of this downburst is the SV that is observed at the upper three measurement levels (Figs. 7 and 8). The SV develops due to the friction by the ground and is elevated to higher elevation by the forward edge of PV (Lundgren et al. 1992). Similar occurrences of SV but in the actual thunderstorm winds were reported in Canepa et al. (2020), Gunter (2019), Chai et al. (2004), and Sherman (1987).

FIG. 7. (a) Proposed outflow structure based on the observations and literature. PV and SV stand for the primary and secondary vortex, respectively. The dashed line separates thunderstorm outflow from the ambient air. (b) Schematics of time histories of the slowly varying mean horizontal ($\bar{u}_H$) and vertical ($\bar{w}$) velocities.

The SV develops opposite vorticity to that in the PV and occurs before the wind directional shift that occurred during the PV (Fig. 8). Reconstructed flow from the tower is analyzed in Fig. 9 using time–space transformation (Sherman 1987). The storm motion u_c that was toward the east and with a constant speed of ~ 20 m s⁻¹ was independently determined from satellite observations (24 m s⁻¹), radar images (20 m s⁻¹), and using Maddox’s (1976) hodograph-based technique (21 m s⁻¹) from the closest sounding. Storm motion was vectorially subtracted from the anemometer measurements to retrieve the storm-relative winds. Notice that the frontal line (dashed curve in Fig. 9a) between the PV and SV experienced frictional retardation close to the surface.

In addition to PV and SV, the reconstructed flow has two different downdrafts passing over the tower. The first downdraft (0238–0239 UTC) had a diameter of ~ 1200 m until it descended to the height of ~ 70 m where it started spreading radially (Fig. 9b). The 30-s mean descending velocity was -0.9 m s⁻¹ at 180 m and a 1-s peak was below -2.5 m s⁻¹. The reflectivity data from De Bilt radar in Figs. 10 and 11 suggest that this downdraft developed in the rare flank of the squall line. The wavelike motion in Fig. 9c that propagated ahead of the PV and SV was reported by Chai et al. (2004) and Knupp (2006) in their lidar and radar measurements of cold outflows. At 180 m, we observe a very regular wavelength of 1100 m and a period of 55 s, which is an order of magnitude shorter than the solitary waves in Knupp (2006) that were observed at heights near and above 1 km. Here, the wavelike motion is also found at 60 m, but the wavelength is slightly longer than at 180 m probably due to surface friction that dissipates more energy than at the higher elevation.

Assuming a linear variation of w with z in the first 3 m, the horizontal component of divergence for an incompressible air is given by

FIG. 8. Instantaneous vertical velocities (w , black lines and the left ordinate) and their slowly varying mean (thick gray line) at all heights. The slowly varying mean of wind direction (α , gray dots) is shown in the right ordinate.

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \approx -\frac{w}{z}. \quad (13)$$

Due to the larger separation between the other three measuring heights and three-dimensional and transient outflow, Eq. (13) might not be a good approximation to other levels. Around 0236 UTC and at 3 m, w averaged over 10 s was -0.45 m s^{-1} (Fig. 8d), producing the horizontal divergence of 0.15 s^{-1} over a 200 m patch. If the area is assumed to be circular, the radial outflow velocity (u_R) from the continuity equation in cylindrical coordinates is given by

$$u_R \approx -\frac{Rw}{2z} = 7.5 \text{ m s}^{-1}, \quad (14)$$

where $R = 100 \text{ m}$ is the radius of the assumed patch. While we do not have other field measurements at 3-m height to confirm this estimate, the sudden change of wind direction around this time (0236 UTC) at higher levels on the main tower (Figs. 8a–c) that was located approximately 200 m south of the 3-m mast (Fig. 1) indicates the existence of a strong radial outflow. Sherman (1987) found similar divergences in a downburst that passed over a tall tower near Brisbane, Australia.

A large region of downward motion with a diameter of $\sim 10 \text{ km}$ is also found between 02:43 and 02:51 at 180 m (Fig. 9a), after

the passage of the squall line (Figs. 10 and 11). While only a small portion of this downdraft penetrated to the ground around 02:48.5 through a narrow region with a diameter of $\sim 500 \text{ m}$, the spread of this air at 60 m and below is observed over a horizontal distance of $\sim 5 \text{ km}$. Also, a strong undercurrent at 3 m is readily observed, except around the PV that contained the strongest radial outflow.

Radar reflectivity in Figs. 10 and 11 resembles a quasi-linear structure, similar to a squall line, that was stretching in the southwest–northeast direction and had an eastward propagation of $\sim 20 \text{ m s}^{-1}$. The higher reflectivity that was observed closer to the forward side of the squall line indicates that the line is more trailing stratiform. The synoptic situation in Fig. 12 shows that the squall line was associated with a low pressure system centered over Scotland. Two occluded fronts in Fig. 12 caused precipitation over the most central and northern Europe (not shown). The observed squall line passed behind the occluded front that was over Germany around 0000 UTC. The eastward propagation of the squall line was in the direction of isohypses of the 500-hPa pressure surface.

The cloud-base height determined from ceilometer measurements was at $\sim 1.5 \text{ km}$ before the downburst arrival (Fig. 13). During the wind event, the cloud base lowered to $\sim 1 \text{ km}$ with large oscillations around this height. The abrupt variations of

FIG. 9. Wind vectors during the passage of downburst over the Cabauw tower using time–space transformation. The 30-s mean values of velocity are plotted every 5 s at the upper three levels and every 10 s at 3 m. Shaded regions are $w < 0$. All vectors are storm-relative and the storm motion was 20 m s^{-1} toward the east. Identified flow structures are additionally highlighted using thick arrows. The dotted line around 0229 UTC shows the estimated position of the frontal line between the PV and SV. Notice that the three subplots are one continuous figure that is split into three rows indicated by the continuation labels.

cloud-base height before the downburst can be caused by virga that is often observed in downburst-producing clouds (Hjelmfelt 1988). The cloud base rose to $\sim 2\text{--}2.5$ km after the thunderstorm passage (partly shown). These differences of cloud-base height before and after the event further resemble a trailing stratiform squall line.

The backscatter coefficient (β) during the downburst ($\sim 0230\text{--}0238$ UTC) also shows some interesting features in Fig. 13. To the author's knowledge, there is no established methodology for the application of ceilometer data in the investigation of the spatiotemporal structure of thunderstorm outflows. While this subject deserves a separate study, the presented analysis is restricted to qualitative statements. We observe closed contours of β during the PV passage, around 0230 UTC, thereby indicating a vortex-like structure in the outflow. The upper region of this structure was above 0.5 km and, therefore, not captured by the anemometers. This observation further suggests that the PV was probably several hundred meters high (Mueller and Carbone 1987). The high values of β are also found around 0236 UTC in the form of a vertical column between surface and cloud base. A small gap

in the column exists at a height of ~ 450 m. Interestingly, this structure occurred around the time of the downdraft occurrence in Fig. 9b.

The sounding from ~ 100 km east of the tower, in Essen, Germany, and about 2.5 h before the event (Fig. 14) shows unidirectional wind shear at all heights with the strongest westerlies occurring around 300–500 hPa. The wind shear in the first 1 km was 0.026 s^{-1} and the strongest upper-level winds exceeded 60 m s^{-1} . The strong winds in the dry layer aloft coupled with saturated and nearly saturated conditions close to the surface indicate an increased likelihood of wet downbursts (Ogura and Liou 1980; Paluch and Breed 1984). This sounding promotes entrainment of dry air into convective downdrafts, which enhances evaporative cooling and increases negative buoyancy of air parcels (Foster 1958; Hookings 1965; Caracena and Maier 1987). The downdraft convective available potential energy (DCAPE), calculated as

$$\text{DCAPE} = -g \int_0^{z_i} \frac{T_d - T_e}{T_e} dz, \quad (15)$$

FIG. 10. Clutter-corrected reflectivity (dBZ) from the De Bilt radar in the Netherlands. The location of Cabauw tower is indicated with the black cross in the third quadrant. Time labels indicate the initial time of scans. All plots show the reflectivity at the lowest elevation of 0.3° . The circular grid resolution is 10 km (dashed lines).

showed a strong dependency on the choice of the initial descending height of the downdraft. Here, T_d and T_e are, respectively, the temperature of the downdraft and environment, and z_i is the initial height of the descending parcel. We integrated Eq. (15) down to the surface ($z = 0$). DCAPE is weak (284 J kg^{-1}) if the downdraft starts descending from the layer of the minimum 100 hPa layer-averaged equivalent potential temperature, θ_e , in the lowest 400 hPa of the sounding. However, if z_i is the lifting condensation level determined from the 600 hPa height instead of the surface, the DCAPE significantly increases to 1560 J kg^{-1} due to the strong entrainment of dry air in the mid- and upper atmosphere (Fig. 14). Further discussion of DCAPE and different choices of z_i is provided in Gilmore and Wicker (1998). In the current sounding, the downdraft is expected to start from a high altitude ($\sim 7 \text{ km}$) and be negatively buoyant down to $\sim 2 \text{ km}$ above ground, as well as in the lowest $\sim 500 \text{ m}$. This analysis suggests that the nocturnal downbursts in European midlatitudes might require a different definition of descending heights in Eq. (15) compared to the U.S. wind

events (Gilmore and Wicker 1998). Also, the predowndraft atmosphere was characterized by low surface-based CAPE (136 J kg^{-1}) and the absence of convective inhibition.

c. Friction velocities and MOST

Noticeable differences between the measured and diagnosed u_* from the MOST are observed during the downburst hour (Fig. 15). The diagnosed u_* is obtained from Eq. (5) without static stability correction:

$$u_* = \frac{\bar{u}_H \kappa}{\ln\left(\frac{z_A}{z_0}\right)}, \quad (16)$$

where z_0 is the climatological value of roughness length (Table 2). Measured u_* values in Figs. 15a–d are determined by applying Eq. (12) to 30-s moving means of eddy covariances calculated from instantaneous data. These 30-s slowly varying means in Figs. 15a–d are then averaged over 20 min to provide the segmentwise mean values in Figs. 15e–h. Traditionally,

FIG. 11. As in Fig. 10, but reflectivity is displayed at the elevation of 10° .

30 s is an insufficient time interval for the estimation of u_* . However, the goal of Figs. 15a–d is to show that the time histories of u_* during the highly transient wind event—such as the investigated downburst—are not transient. At least not as transient as the records of $\bar{u}_H(t)$ (Figs. 5a–d) that were used to diagnose $u_*(t)$.

The discrepancy between measured and diagnosed u_* contradicts the results in Markowski et al. (2019), who reported better agreement between the two. The bias that they observed at higher values of u_* has the opposite trend of ours. However, their work excluded all nocturnal events from the analysis, which is the most likely cause of this disagreement. The results in Fig. 15 are similar to Nieuwstadt's (1978) analysis of Cabauw data. The observed discrepancy between the measured and diagnosed u_* is in part due to the high values of climatologic z_0 in Table 2 (Beljaars and Bosveld 1997). These values might not be suitable for transient thunderstorm winds because the calculated z_0 tends to depend on temperature gradients and stability (Sutton 1953), whereas the actual value only depends on surface characteristics. The discrepancies between the

measured and diagnosed u_* values are the largest after the downburst passed over the tower. The diagnosed values are twice the measured figures. Overall, the difference between the measured and diagnosed values of u_* increases with the height. A spike in the measured u_* around 0230 UTC at 180 m is observed during the PV passage. Given the pronounced $\bar{w} \neq 0$ in the PV, the concept of turbulent flux and u_* is also ill defined according to the classical eddy covariance theory. The overestimation of u_* from MOST implies that the surface shear in the MOST-based numerical models would be much larger than the actual values.

The velocity profiles before, during, and after the downburst are shown in Fig. 16. The fitting of the observed profile with MOST in Figs. 16d–f indicates that the atmosphere was stable during the downburst hour. The stability increases from near-neutral but still stable in the first 20 min to very stable in the last 20 min. Notice that the diagnosed z_0 values from Eq. (16), which were determined using the measured u_* at 3 m, are one order of magnitude smaller than the climatological values provided in Table 2. Although not

FIG. 12. A 500-hPa analysis at 0000 UTC 12 Mar 2008. The dashed lines are contours of 500-hPa surface labeled with heights in meters (source: National Centers for Environmental Prediction reanalysis). The thick lines are the locations of surface fronts (source: Deutscher Wetterdienst analysis). The $\times$ symbol is the location of Cabauw tower.

explicitly shown in Figs. 16d–f, one can infer that the higher z_0 values from Table 2 would significantly diminish the quality of the fit by shifting the MOST curves to the left of the measured velocity profiles. By using more suitable values of z_0 obtained from the measured u_* , however, the MOST managed at reproducing the 20-min means before and after the event. The velocity profiles are not self-similar during the downburst (Fig. 16b) and the deviation from MOST is larger than before and after the downburst

FIG. 13. Ceilometer backscatter (β) contour plot during the downburst hour. The thick line shows the cloud base. The white voids indicate a lack of reliable measurements.

Figure 14

FIG. 14. Sounding of temperature (thick full line) and dew-point temperature (thick dashed line) at 0000 UTC in Essen (Germany), plotted on a skew T – $\log p$ diagram. This sounding represents the environment approximately 2.5 h prior to the downburst passage over the Cabauw tower, and some 100 km east of the tower.

(Fig. 16e). More nose-shape velocity profiles in Figs. 16a–c are found during the downburst passage than in the other two 20-min periods. The deviation of wind profiles from MOST during the 20-min interval of downburst passage should be studied more in the context of wind engineering studies and the application of straight wind tunnels or impinging jet wind chambers in the reconstruction of thunderstorm outflows. Namely, one of the core assumptions in wind tunnel testing is the existence of gradient winds and neutral stratification (Romanic et al. 2020b), but the present study questions both assumptions. Downburst outflows and PV are not driven by gradient winds, but in large by the horizontal pressure gradients that exist along the interface between the cold (dense) outflow and the ambient (light) air.

4. Conclusions

This paper investigated a nocturnal downburst measured on a 213-m tall Cabauw tower in the Netherlands. Ten-hertz anemometer data at 3, 60, 100, and 180 m above ground level (AGL) were used to analyze both mean and turbulent characteristics of this transient and strong wind event that occurred between 0200 and 0300 UTC 12 March 2008. The downburst was characterized by wind speeds that exceeded 30 m s^{-1} at 60 m and above, as well as an abrupt change of wind direction from southwest to west. The shift in wind direction corresponded with slight changes of surface roughness, but, similar to Romanic et al. (2020a), we did not find any significant dependency of flow properties on this change of surface

FIG. 15. (a)–(d) Time series of calculated (full black) and diagnosed (dashed gray) friction velocities (u_*) at four anemometer heights. (e)–(h) Comparison of measured and diagnosed u_* over the three 20-min periods indicated in (a)–(d); see the symbols above (a).

roughness. It also seems that the values of thunderstorm gust factors are not systematically affected by the change of surface roughness, type of the thunderstorm wind and the time of the day when the wind event occurred. More research is needed on these important findings about the dynamics of thunderstorm winds because in other extreme wind environments (e.g., hurricanes), the upstream terrain strongly governs the gust factor distribution and wind flow dynamics.

We identified primary (PV) and secondary (SV) vortex structures at the leading edge of the outflow, as well as the wavelike motion that propagated ahead of the outflow. Observations of a secondary vortex, which is produced due to the interaction of the primary vortex with the surface, are rare in field measurements. The study reconstructed the storm-relative flow field in the outflow using the time–space transformation of anemometer records and the removal of storm motion from velocity measurement. The locations of PV, SV,

and two downdrafts were identified. The estimated diameter of the downdraft downdraft was 1200 m at ~ 70 m AGL. A region of large-scale descent was observed after the thunderstorm passed over the tower. Radar observations and synoptic analysis showed that the downdraft occurred within a squall line that passed over the tower a few hours after the passage of the occluded front. The cloud-base height determined from ceilometer measurements was at ~ 1 km.

The measured velocity profiles were compared against the theoretical predictions of Monin–Obukhov similarity theory (MOST). The MOST profiles based on the values of Obukhov length that correspond to stable stratification provided the best agreement with measurements. With the adjusted value of surface roughness (z_0), MOST properly fitted the velocity profiles before and after the downdraft. However, the theory failed to satisfactorily replicate the wind profiles during the downdraft. This result was not surprising given nose-shape

FIG. 16. (a)–(c) Mean horizontal wind speed observed at the tower during the three 20-min periods indicated in Fig. 15. The profiles are plotted every 30 s. (d)–(f) Nondimensional wind speed as a function of nondimensional height (dots) and their ensemble means (squares). The black curves are MOST predictions for three different Obukhov lengths, L (in m). z_0 values are calculated from Eq. (16) using the measured u_* at 3 m (values provided in each plot).

velocity profiles and the three-dimensionality of thunderstorm outflows. We also compared the measured and MOST-diagnosed friction velocities (u_*) at different heights on the tower. In all cases, MOST without stratification adjustment overestimated the measured u_* . The overshoot was twofold after the downburst passage.

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Data availability statement. Data were available from an FTP server hosted by KHMI.

APPENDIX A

Despiked Velocity Data

Figures A1a–d shows raw data from the ultrasonic anemometers during the downburst passage. The velocity V is calculated as $V = (u_H^2 + w^2)^{1/2}$, where u_H and w are the horizontal and vertical velocities, respectively. The largest number

FIG. A1. (a)–(d) Raw velocity records and (e)–(h) despiked data.

of spikes was detected at 180 m and around the time of peak velocity. Visual inspection of despiked data (Figs. A1e–h) shows the absence of any spurious peaks that “might look” unphysical. However, a more rigorous confirmation of the applicability of the implemented despiking method is provided in Fig. A2, which provides a comparison between the spectra of raw and corrected data. The spikes only affected the highest frequencies by flattening the spectra in that region. This unnatural slope of PSDs in that region shows that the spikes are a random process with uncorrelated frequency content—white noise.

APPENDIX B

Coordinate System Rotation

The transformation of coordinates was necessary for the evaluation of eddy covariances because the anemometer coordinate system is generally not aligned with the mean flow. Figure B1 depicts the relationship between the anemometer coordinate system (x_A, y_A, z_A) and the natural coordinates (x_N, y_N, z_N).

The natural coordinate system is the right-handed system in which the x_N axis is parallel to the mean flow and it increases in the flow direction; therefore, $\bar{v}_N = \bar{w}_N = 0$. After decomposing the velocity record into the mean flow and deviations using Eq. (9), the angles η and θ were expressed as

$$\eta = \text{atan} \left(\frac{\bar{v}_A}{\bar{u}_A} \right), \quad (\text{B1})$$

$$\theta = \text{atan} \left(\frac{\bar{w}_A}{\bar{u}_{h,A}} \right), \quad (\text{B2})$$

where $\bar{u}_{h,A} = (\bar{u}_A^2 + \bar{v}_A^2)^{1/2}$ is the horizontal wind velocity in the anemometer coordinates. The rotation from the anemometer to the natural coordinates is (Tanner and Thurtell 1969)

$$\begin{pmatrix} u_N \\ v_N \\ w_N \end{pmatrix} = \mathbf{R}_\theta \mathbf{R}_\eta \begin{pmatrix} u_A \\ v_A \\ w_A \end{pmatrix}, \quad (\text{B3})$$

with the rotational matrices:

FIG. A2. PSD of reduced turbulence fluctuations of (a)–(d) raw velocities and (e)–(h) despiked data.

$$\mathbf{R}_\eta = \begin{pmatrix} \cos\eta & \sin\eta & 0 \\ -\sin\eta & \cos\eta & 0 \\ 0 & 0 & 1 \end{pmatrix}, \quad (\text{B4})$$

$$\mathbf{R}_\theta = \begin{pmatrix} \cos\theta & 0 & \sin\theta \\ 0 & 1 & 0 \\ -\sin\theta & 0 & \cos\theta \end{pmatrix}. \quad (\text{B5})$$

After introducing $\text{CE} = \cos\eta$, $\text{SE} = \sin\eta$, $\text{CT} = \cos\theta$ and $\text{ST} = \sin\theta$, we expand Eq. (B3) in terms of mean and fluctuating velocity components:

$$\bar{u}_N = \bar{u}_A(\text{CT})(\text{CE}) + \bar{v}_A(\text{CT})(\text{SE}) + \bar{w}_A(\text{ST}), \quad (\text{B6})$$

$$\bar{v}_N = \bar{w}_N = 0, \quad (\text{B7})$$

$$u'_N = u'_A(\text{CT})(\text{CE}) + v'_A(\text{CT})(\text{SE}) + w'_A(\text{ST}), \quad (\text{B8})$$

$$v'_N = v'_A(\text{CE}) - u'_A(\text{SE}), \quad (\text{B9})$$

$$w'_N = w'_A(\text{CT}) - u'_A(\text{ST})(\text{CE}) - w'_A(\text{ST})(\text{SE}). \quad (\text{B10})$$

The expressions (B6)–(B10) are used to derive the transformation of momentum fluxes:

$$\begin{aligned} \overline{u'_N u'_N} &= \overline{u'_A u'_A}(\text{CT})^2(\text{CE})^2 + \overline{v'_A v'_A}(\text{CT})^2(\text{SE})^2 + \overline{w'_A w'_A}(\text{ST})^2 \\ &\quad + 2\overline{u'_A v'_A}(\text{CT})^2(\text{CE})(\text{SE}) + 2\overline{u'_A w'_A}(\text{CT})(\text{ST})(\text{CE}) \\ &\quad + 2\overline{v'_A w'_A}(\text{CT})(\text{ST})(\text{SE}), \end{aligned} \quad (\text{B11})$$

$$\overline{v'_N v'_N} = \overline{v'_A v'_A}(\text{CE})^2 + \overline{u'_A u'_A}(\text{SE})^2 - 2\overline{u'_A v'_A}(\text{CE})(\text{SE}), \quad (\text{B12})$$

$$\begin{aligned} \overline{w'_N w'_N} &= \overline{w'_A w'_A}(\text{CT})^2 + \overline{u'_A u'_A}(\text{ST})^2(\text{CE})^2 \\ &\quad + \overline{v'_A v'_A}(\text{ST})^2(\text{SE})^2 - 2\overline{u'_A w'_A}(\text{CT})(\text{ST})(\text{CE}) \\ &\quad - 2\overline{w'_A v'_A}(\text{CT})(\text{ST})(\text{SE}) + 2\overline{u'_A v'_A}(\text{ST})^2(\text{CE})(\text{SE}), \end{aligned} \quad (\text{B13})$$

FIG. B1. Rotation from the anemometer coordinate system (x_A , y_A , z_A) through the angles η and θ to the natural coordinates (x_N , y_N , z_N). The horizontal plane in the anemometer frame of reference is shaded. Also, $(\bar{u}_A, \bar{v}_A, \bar{w}_A)$ are the mean velocity components in the anemometer coordinate system and $\bar{\mathbf{V}}$ is the mean velocity vector.

$$\begin{aligned} \overline{u'_N w'_N} &= \overline{u'_A w'_A} (CE) [(CT)^2 - (ST)^2] - 2\overline{u'_A v'_A} (CT)(ST)(CE)(SE) \\ &\quad + \overline{w'_A v'_A} (SE) [(CT)^2 - (ST)^2] - \overline{u'_A u'_A} (CT)(ST)(CE)^2 \\ &\quad - \overline{v'_A v'_A} (CT)(ST)(SE)^2 + \overline{w'_A w'_A} (CT)(ST), \end{aligned} \quad (B14)$$

$$\begin{aligned} \overline{u'_N v'_N} &= \overline{u'_A v'_A} (CT) [(CE)^2 - (SE)^2] + \overline{w'_A v'_A} (ST)(CE) \\ &\quad - \overline{u'_A w'_A} (ST)(SE) - \overline{u'_A u'_A} (CT)(CE)(SE) \\ &\quad + \overline{v'_A v'_A} (CT)(CE)(SE), \end{aligned} \quad (B15)$$

$$\begin{aligned} \overline{v'_N w'_N} &= \overline{v'_A w'_A} (CT)(CE) - \overline{u'_A w'_A} (CT)(SE) \\ &\quad - \overline{u'_A v'_A} (ST) [(CE)^2 - (SE)^2] + \overline{u'_A u'_A} (ST)(CE)(SE) \\ &\quad - \overline{v'_A v'_A} (ST)(CE)(SE). \end{aligned} \quad (B16)$$

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