Cold pools over the Netherlands: A statistical study from tower and radar observations

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Abstract

We provide a detailed analysis of convectively generated cold pools (CPs) over flat midlatitude land, combining ten-year high-frequency time series of measurements at several heights available from the 213-m tower observatory at Cabauw, the Netherlands, with a collocated 2D radar rainfall dataset. This combination of data allows us to relate observations of the CP's temporal and vertical structure to the properties of each CP's parent rain cell, which we identify by rain-cell tracking. Using a new detection method, based on the anomalies of both the vertically averaged wind and the temperature, we monitor the arrival and passing of 189 CPs during ten summers (2010-2019). The time series show a clear signature of vortex-like motion along the leading CP edge in the vertical and horizontal wind measurements. The arrival of CP gust fronts is characterized by a steep decrease in both temperature and moisture, with a recovery time of approximately two hours. We see no evidence of moisture rings on the gust front edge, and therefore no indications for thermodynamic convective triggering. From the tower data, we obtain a median CP temperature drop of $T_{\rm drop} \approx -2.9$ K and a height-averaged horizontal wind anomaly of $\Delta u_{\rm max} \approx 4.4 \,{\rm m} \cdot {\rm s}^{-1}$. Relating the individual CP's horizontal wind anomalies and temperature drops, we confirm the validity of the theoretical density current relationship, $\Delta u_{\text{max}} \propto T_{\text{drop}}^{1/2}$. We further propose a simple statistical model to relate the CP strength defined by T_{drop} to the environmental properties influencing the CP: rain intensity and lower boundary-layer saturation. A multivariate linear regression suggests a 1 K colder CP for a 4 mm·hr⁻¹ more intense rain cell (instantaneous area-averaged rain intensity) or for a 2.5 K larger pre-CP dew-point depression.

KEYWORDS

atmospheric convection, cold pools, convective downdrafts

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1 | INTRODUCTION

A convectively generated cold pool (CP) is a subcloud volume of air, which is cooled by the partial evaporation of precipitation. This dense, cold air descends, as it is negatively buoyant relative to its surroundings, and is often further accelerated by the drag exerted by falling hydrometeors (Wakimoto, 2001). When hitting the ground, the CP spreads along the surface away from its source as a density current (Charba, 1974), introducing cold dense air underneath warmer environmental air. Thereby, vortical motion builds up, which is measurable as the combination of horizontal wind, distinctive horizontal convergence lines, and vertical mass fluxes-often referred to as the "CP outflow boundary" or "CP gust front". The amount of evaporation, and hence cooling within a CP, depends on the rain intensity and area of the generating rain cell, as well as the environmental atmospheric profiles of temperature and relative humidity. The coldest outflows thus stem from high-based thunderstorms that precipitate into very dry boundary layers (Markowski and Richardson, 2010). Beyond these macrophysical conditions, microphysical parameters, crucially the drop-size distribution, influence the rain evaporation (Seifert, 2008).

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CP characteristics are often studied using numerical simulations that now approach the fine scales needed to resolve some CP properties, that is, horizontal grid resolutions of substantially less than 1 km (Drager and van den Heever, 2017; Fournier and Haerter, 2019; Cafaro and Rooney, 2018; Drager et al., 2020; Meyer and Haerter, 2020). Observational work is, however, indispensable as a means of comparison and validation for numerical studies. Oceanic measurement campaigns have provided insight into the dynamics of ensembles of CPs over the tropical and subtropical oceans (Young et al., 1995; Zuidema et al., 2012; Vogel, 2014; de Szoeke et al., 2017; Chandra et al., 2018). As in simulations, CPs observed over the tropical oceans show relatively weak temperature anomalies of typically -1 to -1.5 K and wind gusts of 2-2.5 m·s⁻¹. An important morphological feature arising in simulated oceanic CPs are so-called moisture rings, that is, bands of enhanced water vapor, which build up near the gust front as the CP spreads. Studies over ocean find minor increases in moisture, $0.25 \text{ g} \cdot \text{kg}^{-1}$ (de Szoeke *et al.*, 2017), measured ahead of the gust front, whereas subsequent decreases in moisture, measured behind the gust front, are nearly an order of magnitude larger. CPs over land vary much more strongly than those over ocean and can reach temperature anomalies as deep as -17 K and wind gusts larger than $15 \text{ m} \cdot \text{s}^{-1}$ (Engerer *et al.*, 2008). Observational studies of CPs over land are often focused on case studies and are mostly based in the continental United States (Mueller and Carbone, 1987; Wakimoto, 1982; Engerer et al., 2008; Hitchcock *et al.*, 2019; van den Heever *et al.*, 2021). In contrast to oceanic CPs, the moisture signal is not always prevalent in observations over land: a recent statistical analysis of midlatitude CPs over Hamburg, Germany, finds no strong signature of moisture rings near the gust front, but the authors do find a pronounced, longer-lasting moisture signal that increases up to one hour after the first detection of the CP (Kirsch *et al.*, 2021). Simulating continental CPs and detecting their edges through a buoyancy anomaly method, Drager *et al.* (2020) conclude that moisture rings over land may only occur in conditions of moist surfaces, but not for drier conditions.

A CP can interact with its environment and other CPs, producing updrafts and potentially triggering new convective clouds (Purdom, 1976; Weaver and Nelson, 1982; Droegemeier and Wilhelmson, 1985a; 1985b; Wilson and Schreiber, 1986; Moncrieff and Liu, 1999; Tompkins, 2001b; Torri et al., 2015). The triggering of new convection can happen through mechanical forcing, due to the lifting of air along the gust front of the CP, or thermodynamic forcing, due to the abovementioned moisture rings that provide additional buoyancy, favoring new convection at the edge of the CP (Tompkins, 2001a; Torri et al., 2015; Drager et al., 2020). An incessant sequence can unfold, where the new precipitating clouds themselves may create new CPs that in turn can result in a new precipitation event, which then produces a new CP, etc. CPs may therefore be a key ingredient to the understanding of how clouds organize spatially and temporally into larger-scale precipitating systems (Böing, 2016; Haerter, 2019; Haerter et al., 2019). The "upscale communication" between small-scale individual CPs (~ 10 km horizontally and ~ 1 hr temporally) and the larger-scale spatial organization of the cloud field (\geq 100 km, respectively \geq 3 h, such as mesoscale convective systems) has been explored in both theoretical (Rotunno et al., 1988; Jeevanjee and Romps, 2013; Haerter et al., 2019; Haerter et al., 2020) and observational studies (Zipser, 1977; Feng et al., 2015; Zuidema et al., 2017). All the more, there is a need for better understanding of the processes that determine the temporal evolution of CP properties, such as their spatial extent, lifetime, and strength (Drager and van den Heever, 2017). Defining the relationship between the properties of a given CP and the precipitation cell that produces this CP (termed the "parent" rain cell) can serve as a useful benchmark for numerical simulations.

In this study, we provide an observation-based analysis of CPs over midlatitude coastal land, relating CP properties to their parent rain cell and the environment. We use weather measurements from the Netherlands, a region with climate that is a hybrid between a moist oceanic and a drier continental regime. It is thus interesting to study CPs in such a region, because CPs are strongly affected

by boundary-layer moisture. We combine boundary-layer measurement tower data and radar data and further develop a method to detect CPs from tower measurements based on a previous study by de Szoeke et al. (2017). The measured horizontal wind and temperature at the time of CP detection allow us to empirically test the theoretical scaling between CP propagation speed and temperature depression (von Karman, 1940; Benjamin, 1968). Radar imagery is then utilized to visualize and track the rain cells in the tower's surroundings. Using the distance of the rain cells to the measurement tower and the wind direction during CP passage, we attribute a specific parent rain cell to each CP. Combining the point measurements from the tower with the tracked rain cells allows us to analyse the relation between CP temperature anomaly, the pre-event saturation of the boundary layer, and the rain intensity of the CP's parent rain cell.

2 | DATA AND METHODS

2.1 | Data

We use data from a 213-m boundary-layer measurement tower at the Cabauw Experimental Site for Atmospheric Research (CESAR observatory), located in Cabauw (51.971°N, 4.927°E), the Netherlands. These data enable us to study the temporal evolution of lower boundary-layer properties before, during, and after the passage of a CP gust front over land, in a temperate maritime climate. We focus on the summer period (May-September), to capture the maximum convective activity in the region. The time series used consist of 1-min averaged measurements of temperature, dew-point temperature, horizontal wind speed, and wind direction at six different heights above ground (10, 20, 40, 80, 140, 200 m) for the period 2010-2019. Temperature is measured with Pt500 elements, placed in unventilated screens to minimize the influence of radiation and precipitation; dew-point temperature is measured with EplusE 33 polymer-based relative humidity sensors, which are heated to decrease measurement problems during humid conditions; horizontal wind speed and wind direction are measured with a cup-anemometer and vane combination that rotate with the direction of the wind and record the velocity of the propellers (Bosveld et al., 2020). We additionally use 0.1-s records of water vapor concentration and vertical wind speed measured at heights of 60 and 180 m that are available only for the summer of 2019, measured with a Gill-R50 sonic anemometer (Bosveld et al., 2020). Tower data were retrieved from the CESAR Data Portal (CESAR).

Additionally, we use a rainfall radar dataset from the Royal Netherlands Meteorological Institute (KNMI) to track and study the rain events generating the CPs for the period 2010–2019. These 2D horizontal data are composites of radar reflectivities from both of the KNMI weather radars, Den Helder and Herwijnen. The resultant dataset encompasses five-minute precipitation heights on a 1 km \times 1 km grid, which have been adjusted employing validated and complete rain-gauge data from the KNMI rain gauge networks, and provides data for the entire land surface of the Netherlands.

To look for visible gust fronts, we further use 5-min time resolution radar imagery from the Herwijnen C-band polarimetric Doppler radar tower (51.837°N, 5.138°E). The Herwijnen radar scans the surrounding atmosphere at 15 different elevation angles, starting from the horizontal and upwards. The finest radial resolution of the Herwijnen radar is approximately 225 m and the azimuthal resolution is approximately 1 degree.

All the radar data were retrieved from the open-access KNMI Data Portal (KMMI, 2020).

2.2 | Algorithm for the detection of cold pools from tower measurements

We build an algorithm to detect CPs from tower measurements, based on a temperature detection algorithm used previously for oceanic CPs (de Szoeke *et al.*, 2017). We modify this algorithm by tailoring the temperature threshold to continental conditions, imposing a time constraint on the temperature anomaly, and adding a criterion on the horizontal wind anomaly. This additional criterion is imposed to ensure the existence of a wind gust along with the detection of cold air, thus incorporating the two main characteristics of a CP. For any given day, we use the 1-min time series of temperature at the 10-m tower level, and 1-min time series of wind speed at the six tower levels (10, 20, 40, 80, 140, and 200 m).

2.2.1 | Temperature criterion

The algorithm by de Szoeke *et al.* (2017) identifies CP gust fronts from 1-min surface temperature over a tropical oceanic surface. It is "designed to be sensitive to asymmetric cooling events visible in the time series, yet insensitive to high-frequency noise, to exclude false positives". We use this algorithm as a first step in CP detection. We first smooth the temperature time series with a running 11-min centered window. A series of threshold operations is then applied to this smoothed time series (with a resolution of 1 min), to identify and "record" CPs along with their properties (Figure 1a).

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FIGURE 1 Exemplifying cold pool detection by two-step criterion. (a) Daily time series of temperature, measured at the 10 m level at 1-min temporal resolution and smoothed with an 11-min centered window. The red symbols and two horizontal lines mark the initial temperature drop δt of a detected CP event. (b) Horizontal wind anomaly averaged over all tower heights (Section 2.2) measured at 1-min temporal resolution. The anomaly is computed with respect to a 2-hr centered running temporal average. The horizontal dashed blue line indicates four standard deviations from the daily mean horizontal wind anomaly, exceedance of which is used as a criterion for the detection of strong wind anomalies. (c) Sketch of a cold pool (blue shaded area) crossing the Cabauw tower. The red arrows indicate the propagation velocity and internal circulation of the cold pool, together composing the measured horizontal wind anomaly. The levels of temperature and horizontal wind measurements (10, 20, 40, 80, 140, and 200 m) are indicated by solid horizontal black lines [Colour figure can be viewed at wileyonlinelibrary.com]

- 1. The temperature at a given time step is defined as a minimum temperature if it is lower than any temperature within the preceding 20-min time window.
- 2. CP events are defined by combining multiple such temperature minima if they are consecutive (i.e., separated by 1 min), or if lying within 20 min of each other given that the temperature during that time window does not exceed either of them by 0.5 K. This way we allow for small temperature fluctuations within the CP interior without detecting it as two separate CP events. We choose these values following the algorithm proposed in de Szoeke *et al.* (2017). The temperatures at the beginning and end of a CP event are denoted as the first and last T_{min} .
- 3. For each detected CP event, the temperature drop δT is defined as the absolute difference between the last T_{\min} and the maximum smoothed temperature in the 10-min time window preceding the first T_{\min} (Figure 1a). A time interval Δt is defined as the time elapsed between the first and the last T_{\min} .
- 4. A CP event is recorded if δT exceeds a 1.5-K threshold and the time interval Δt does not exceed 60 min. The temperature threshold is raised compared with the one used in de Szoeke *et al.*, 2017 (aimed at detecting CPs over a tropical oceanic surface) to reduce the

signal-to-noise ratio, caused by the higher temperature fluctuations over land compared with the (tropical) ocean.

5. A refined temperature drop T_{drop} is defined as the difference between the maximum unfiltered temperature within 10 min preceding the first T_{min} and the minimum unfiltered temperature within the temperature drop. We use this stronger temperature drop to have a more accurate measure of the effective cooling due to the CP. We note that, whereas the δT values are systematically too low due to smoothing, they may at times be very high if they include local fluctuations.

2.2.2 | Wind criterion

A further criterion is then added to the detection algorithm, as a novelty with respect to the algorithm used by de Szoeke *et al.* (2017), to ensure that there is a wind gust associated with each detected temperature decrease. For this purpose we use the time series of horizontal wind speed at the six tower levels. For each day we smooth the 1-min time series with a running 2-hr centered window and subtract the smoothed time series from the original time series to obtain the "horizontal wind anomaly". We

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FIGURE 2 Schematic of rain cell attribution algorithm. (a) Overview of radar data domain with the location of the Cabauw tower marked. The black contour line marks the political boundary of the Netherlands. (b) Schematic of geometrical requirements on the rain cell position during $t \in [t_0 - \delta \min(t_0, t_{\min})]$ based on the wind direction of the detected CP gust front at time t_0 . Only rain cells within the yellow half-circle are accepted. (c) Schematic of requirement on temporal overlap of rain cells that exist in the time window $t \in [t_{\min}, t_{\max}]$ [Colour figure can be viewed at wileyonlinelibrary.com]

choose the 2-hr smoothing window to extract short-term fluctuations from long-term wind variability. To extract the vertically coherent signals in the data set, we compute the average wind speed over all six tower heights. This way, we reduce random (or turbulent) fluctuations, and a wind gust visible at all six heights will rise above the noise. We call this variable the "height-averaged horizontal wind anomaly" (Δu). For simplicity, we team the maximum of this variable in a given time window the "wind peak" Δu_{max} .

The detection algorithm then scans the events recorded by the temperature criterion, and ultimately saves an event as a "cold pool" if there is a wind peak within 10 min preceding the first T_{min} and the time of the last T_{min} that exceeds four standard deviations (4σ) of the daily 1-min Δu time series. An example, illustrating the algorithm (Figure 1), shows the 10-m temperature time series and the height-averaged horizontal wind anomaly time series of a specific day where a CP was detected (August 27, 2019).

2.2.3 | Parameter sensitivity

The parameters for the algorithm were initially tuned based on two case studies, where the front of the CPs was clearly visible in the radar images due to dust and/or insects drafted up in the convergence zone (Herwijnen radar images, May 29, 2018, 1430–1530 UTC and August 27, 2019, 1630–1700 UTC (Figure 3), presented in Kruse, 2020). This allowed us to visually determine the time instance at which the CP front should be detected at the measurement tower, revealing typical CP signals in the time series that we should look for and against which the algorithm was calibrated. The thresholds on temperature and wind peak were chosen to capture these CP cases from the daily temperature and wind time series, and were kept as high as possible in order to find similar cases of strong, clear CP signals throughout the year and to exclude sea breezes. Reducing the threshold in temperature anomaly from 1.5 K to 1 K adds only a few extra cases (an additional 15% for the summer of 2019). By contrast, reducing the threshold on the wind speed from 4σ to 3σ nearly doubles the number of cases detected. However, the additional cases are of very similar nature, that is, the gust fronts are detected shortly prior to the presence of a rain cell over the tower, with the only difference that the rain cells are weaker. This means that the additional CPs detected are generated by rain cells with intensities that are often lower than the threshold of 1 mm·hr⁻¹ set for the detection of convective rain cells (Section 2.3).

2.3 | Attribution of a rain cell from radar data

2.3.1 | Rain cell tracking

We use an Iterative Raincell Tracking (IRT) method (Moseley et al., 2014; 2019) to track rain cells in time and space from the gridded radar rainfall product. The IRT locates spatially contiguous areas of rainfall, termed objects, and tracks them in time if they overlap with objects in subsequent time steps. This allows the definition of rain cell *tracks*, extending over a time window $t \in [t_{\min}, t_{\max}]$. For the object identification, a threshold of 0.08 mm per 5 min, corresponding to $I \approx 1 \text{ mm} \cdot \text{hr}^{-1}$, is imposed along with a requirement of a minimum of four contiguous rainy pixels (one pixel corresponds to an area of approximately 1 km \times 1 km). To each detected CP a rain cell track is then attributed based on a multistep algorithm (Figure 2). First, we select all rain tracks that exist for at least $\Delta = 10 \text{ min}$ during the time interval $\delta = 30$ min preceding the CP detection time t_0 at any point in the domain. The tracks are allowed to end before the CP is detected, defining a "rain cell time step" $t_{\text{RC}} = \min(t_0, t_{\text{max}}) \in [t_0 - \delta + \Delta, t_0]$. In a second step, all rain tracks having a closest edge further than $r_{\text{max}} = v_{\text{max}} \cdot \delta = 18$ km away from the location of the Cabauw tower (\mathbf{x}_{Cabauw}) at time step t_{RC} are discarded. The choice of this radius is based on the assumption of an upper bound on CP propagation speed $v_{max} = 10 \text{ m} \cdot \text{s}^{-1}$, implying that gust fronts generated by rain that falls further away cannot reach the tower in time. In a final step, the domain is cut in half-planes based on the wind direction measured at the tower during CP passage (ϕ_{CP}): all rain cells with centers of mass (COM) located in a direction relative to Cabauw tower that deviates by more than 90° from $\phi_{\rm CP}$ are discarded (Figure 2). In the majority of cases, this three-step process allows for the identification of a unique rain cell, which is assumed to be the parent rain cell of the gust front measured in the tower time series. We point out that the number of CPs with a unique attributable rain cell is 116 out of 189, meaning that the statistics involving CPs in connection to the rain have fewer data points. We note that, in general, the detected CPs are located close to the edge of their parent rain cell. This is quantified by the relative distance of the RC's COM to Cabauw tower, divided by the approximate radius of the RC when assuming circular shape, both taken at $t_{\rm RC}$:

$$d = \frac{|\mathbf{x}_{\text{RC, COM}} - \mathbf{x}_{\text{Cabauw}}|}{\sqrt{A_{\text{RC}}/\pi}},\tag{1}$$

where $\mathbf{x}_{\text{RC,COM}}$ is the position of the rain cell's center of mass and A_{RC} is the area of the rain cell. We find that *d* is distributed around a mean value of $\langle d \rangle = 1.7$ with a tail reaching values as large as d = 5 (Figure A1), where d = 1 corresponds to the CP gust front being detected directly at the edge of the rain cell. The relatively large number of cases with d < 1 is an artefact of the definition of *d* being based on the assumption of circular RCs. In reality, many CPs are generated by RCs of elongated shape that are oriented perpendicular to the vector $\mathbf{x}_{\text{RC, COM}} - \mathbf{x}_{\text{Cabauw}}$ (Figure A2).

2.3.2 | Qualitative analysis of weather situations

The combination of radar data with the CP detection from tower measurements allows the observation of the large-scale weather situation in which the CPs live. As expected, the algorithm detects gust fronts in very diverse weather situations, which we identify qualitatively as isolated convection (Figure A2a), elongated precipitation cells resembling squall lines (Figure A2b), large-scale fronts (Figure A2c), and mesoscale convective systems, characterized by low large-scale wind (Figure A2d). The different weather situations that the CPs are nested in are each defined by a particular boundary-layer wind shear, background wind direction with respect to the direction of propagation towards the tower, soil moisture, etc. These factors contribute to the spread in the measurements of CP properties (Section 3).

3 | RESULTS

3.1 | Cold-pool structure

$3.1.1 \mid \text{Case study}$

We first discuss an individual CP event as seen in C-band radar imagery. In the radar reflectivity, measured at the Herwijnen radar tower (Figure 3, left panels), one can clearly see the development of a quasicircular gust front (light green shades), the CP edge, spreading around an area of high reflectivity (dark blue) that we associate with the rain event. The gust front is visible in the reflectivity due to insects and/or dust carried aloft by the convergence of air at the CP boundary (Markowski and Richardson, 2010). In the radial Doppler velocity measurements (Figure 3, right panels), one can see that the area delimited by the gust front is characterized by outward movement (positive radial velocities ranging from $5-10 \text{ m} \cdot \text{s}^{-1}$). If we approximate the spreading gust front with a temporally growing circle of radius r(t), we obtain a near-constant horizontal propagation speed $v(r) \equiv dr(t)/dt \approx 7 \,\mathrm{m \cdot s^{-1}}$. We note that this is a unique case, as we have a strong CP signal that spreads almost perfectly around the radar tower-this is generally not the case in the rest of our data set. Interestingly, the wind peak measured at Cabauw tower at the time when the radar gust front seems to pass by the tower actually comes from the direction of the smaller rain cell, located southwest of the tower. We use this example to stress the added benefit of verifying the wind direction of the wind gust when attributing a generating rain cell to a CP signal.

3.1.2 | CP composite time series

The composites for horizontal wind anomaly and temperature anomaly (Figure 4) are drawn from 189 CPs detected in the 10 summers from 2010–2019, while the composites for vertical wind and moisture anomaly (Figure 5) are drawn from only 18 CPs detected in the summer of 2019. The discrepancy in CP count is due to the availability of data from different measurement instruments (Section 2.1). Most CPs are detected in the

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afternoon, coinciding with the time of day when convection is most active over land (Figure A3).

We compute the composites by first centering the individual CP time series on their respective times of maximum horizontal wind anomaly (t_0) , and retain the data for 60 min before and 120 min after t_0 , resulting in a set of time series which each have the same number of time steps. For each time step, we then average over all time series, yielding the mean time series. To compute the composites of the anomalies x'(t) of a quantity x(t), such as horizontal wind, temperature, and water vapor concentration, we first remove the respective pre-CP temporal mean from the time series, that is,

$$x'(t) \equiv x(t) - \overline{x} , \qquad (2)$$

where \bar{x} is the time average over the 51 1-min time steps from $t_0 - 60$ to $t_0 - 10$ min, hence the time window preceding the arrival of the CP. The 10-min margin was chosen to ensure that the CP signal does not influence the mean. Furthermore, for the temperature anomaly, we remove the effects of the diurnal cycle by subtracting the two-hour running mean. We verified that the one-year composites of temperature and wind, although they are somewhat more noisy, are comparable with the ten-year composites of the same variables. This makes us confident that one year of data is representative of a larger data set.

The edge of the composite CP is characterized by a strong positive horizontal wind anomaly (gust front) seen at all tower heights (Figure 4a). Before and after passage of the gust front, the wind anomaly increases monotonically with the tower height, as one would expect in the surface layer (see Figure 4d). By contrast, within a window of approximately six minutes enclosing $t = t_0$, the largest value of horizontal wind anomaly is measured at an intermediate tower level, near z = 80 m (Figure 4c). Since the measured horizontal wind anomaly corresponds to the superposition of the propagation speed of the CP front and the internal CP circulation (Rooney, 2018), we interpret this window as the time interval where the vortical circulation within the CP head affects the measured horizontal wind speeds. If we assume the average CP gust front to be propagating at $u \approx 0.67 u'(t_0)_{10m} = 2.7 \text{ m} \cdot \text{s}^{-1}$, following the relation found in Goff (1976), this would imply that the width of the CP head is approximately 1 km, considering the transit time of six minutes.

The horizontal wind anomaly is preceded by a negative temperature anomaly that occurs simultaneously and at the same rate at all tower heights. After t_0 , the lower heights *z* show systematically deeper anomalies T'(z). We 8



FIGURE 4 Composite time series of (a) horizontal wind anomaly u' and (b) temperature anomaly T'. These composites include 189 CPs detected in the summers 2010–2019. Blue lines show the ensemble mean; blue shaded areas show the standard deviation, computed from the CP ensemble, indicating the ensemble spread between the different CPs at each given time, and computed at the height of the strongest signal for each variable (80 m for the horizontal wind, 10 m for the temperature). Vertical lines highlight times $\Delta t = 0 \min$ (red) and $\Delta t = 10 \min$ (blue). Insets show (c) u' at the heights measured at the tower at $\Delta t = 0 \min$; (d) analogous to (c) but at $\Delta t = 10 \min$; (e) T' at the heights measured at the tower at $\Delta t = 10 \min$; (d) analogous to (c) but at $\Delta t = 10 \min$; (e) for an expression of the tower at $\Delta t = 10 \min$, with linear fit used to estimate CP height (intercept: 503.8, slope: -304.7) [Colour figure can be viewed at wileyonlinelibrary.com]

interpolate T'(z) linearly at 10 min towards T'(z) = 0, to obtain a rough estimate of the height z_0 where the temperature anomaly disappears (Figure 4b, inset). We estimate $z_0 \approx 500$ m as an indication for a height scale for the body of the composite CP. We note that the temperature anomaly is fully recovered at all measurement heights after approximately two hours from the beginning of the temperature drop.

The circulation within the CP head is further characterized by the vertical wind peak (updraft) that precedes the horizontal wind anomaly, as reflected in a positive vertical wind anomaly 1–2 min before t_0 , which exceeds four standard deviations of the fluctuations in the time series (Figure 5a). Here, the standard deviation represents the fluctuation of the composite vertical velocity time series in the time window [$t_0 - 60, t_0 + 120$ min], so, although the vertical wind is noisy, the exceedance of this line indicates a clear signal of a strong updraft. The updraft signal is strongest at the highest measurement level (180 m).

The water vapor concentration starts decreasing at the same time as the vertical wind peak occurs at both measurement heights, confirming that the CP interior is dry. There is not a clear signal of enhanced moisture before t_0 , indicating that moisture rings may not be an evident characteristic of the CPs in this study. We note, however, that the water-vapor concentration starts decreasing four to five minutes after the temperature has started decreasing, meaning that the CP head is more moist than the body. The moisture anomaly is recovered within one to two hours. Whereas all previously discussed composite characteristics are mostly in line with the findings from Kirsch et al. (2021) for CPs over Hamburg, Germany, the moisture signal differs significantly, showing moistening rather than drying in the interior of the CPs. The absence of a wind criterion with a high threshold may lead to the inclusion of CPs measured at different points of their lifetime, or from different types of rain events, which would affect the moisture signal. Furthermore, Drager et al. (2020) show in their simulations that the moisture content in the interior and ahead of CP fronts depends crucially on the soil moisture: over dry soils, their CPs show an increase in moisture, similar to the observations in Kirsch et al. (2021), whereas for wet soils the moisture signal shows the same characteristics as our measurements with dry air in the interior, but with the addition of moisture rings ahead of the CP front.

In Figure 6 we provide a sketched summary of the observed CP characteristics. The circulation, temperature,



FIGURE 5 Composite time series of (a) vertical velocity and (b) water-vapor concentration (WVC) anomaly. These composites include 18 CPs detected in the summer of 2019. Blue shaded areas in each panel show the standard deviation, computed from the CP ensembles, indicating the ensemble spread between the different CPs at a given time, and computed at the height of the strongest signal for each variable (180 m for the vertical velocity, 180 m for the WVC). The vertical red line highlights time $\Delta t = 0 \min$ [Colour figure can be viewed at wileyonlinelibrary.com]

and moisture signal indicate that the CP edge (measured at $t_0 - 5$ min) is characterized by a moist, cold, updraft; the CP head (measured at t_0) is characterized by cold, dry air and increased vorticity; and the body of the CP (measured at $t_0 + 10 \text{ min}$) is characterized by dry air, with the largest temperature anomalies at the surface. The thermodynamic structure of the CP interior is characterized by increased atmospheric stability of approximately 5 K·km⁻¹, as can be seen in the stratification of the temperature anomaly (Figure 4b), where the lowest level shows the largest cooling. This stratification does not exist in the moisture signal, which appears to be homogeneous drying through the CP's height, since the anomalies at 60 and 180 m are similar in value (Figure 5b). The recovery time for temperature and moisture after the passage of the CPs seems to vary strongly among CPs, as indicated by the ensemble variance in temperature and moisture anomaly (blue shading in Figures 4 and 5).

3.2 | CP strength

The "strength" of a CP can be characterized dynamically, by its propagation speed, and thermodynamically, by its temperature anomaly. Here we wish to quantify the strength of an ensemble of CPs. Early studies show that, for incompressible, inviscid, and irrotational (i.e., no internal motion) density currents in unstratified flows, the propagation speed u can be related to the relative density difference between the interior of the density current and its surrounding environment (von Karman, 1940; Benjamin, 1968). Considering that the relative density difference can be approximated with the relative temperature difference, a general equation describing the relationship between the propagation speed u and the temperature anomaly ΔT is

$$u = k \sqrt{gH \ \frac{\Delta T}{T_0}},\tag{3}$$

where *k* is the "internal Froude number" *k* (Benjamin, 1968; Wakimoto, 2001), *g* is the gravitational acceleration, *H* the CP height, ΔT the temperature difference between the CP and its environment, and T_0 the air temperature of the environment. Since CPs are density currents in a nonidealized environment, they are exposed to dissipation effects, such as surface friction and turbulent mixing, which are usually included within *k*. The inviscid case hereby represents a special case with $k = \sqrt{2}$, while meteorological studies have found values $k \approx 0.7$ to be more realistic (Wakimoto, 1982; 2001; Markowski and Richardson, 2010). We test the above relation here, assuming a Froude number of k = 0.7.



FIGURE 6 Summarizing sketch of measured CP properties. On the left is an approximate sketch of which part of the CP is being measured at the tower at a given point in time: (a) at $t = t_0 - 5$ min, the edge of the CP is being measured at the tower; (b) at $t = t_0$, the head of the CP; (c) at $t = t_0 + 10$ min, the body of the CP. On the right, a depiction of the measured CP properties corresponding to each part of the CP measured at its lowest 200 m is given: the edge of the CP shows moist, well-mixed air, along with an updraft; the head of the CP shows dry, colder, well-mixed air, and the signature of a vortex ring is reflected in the horizontal wind anomaly; the body of the CP shows dry air and a layered temperature anomaly, with the largest cold anomaly within the bottom layers [Colour figure can be viewed at wileyonline]

For each CP, we estimate the environmental temperature T_0 from the temperature at 10 m, averaged over the time window $[t_0 - 60, t_0 - 10 \text{ min}]$, as done previously. A scatter plot of the wind peak Δu_{max} against the relative temperature drop $T_{\rm drop}/T_0$ for the 189 CPs detected in the ten summers 2010-2019 (Figure 7) indeed suggests increasing gust front speed for larger temperature drops. Viewing Δu_{max} as a proxy for the CP's total kinetic energy density and T_{drop} as a representation of the CP's potential energy density, this indicates a monotonic relation between the kinetic and potential energy of the CP (Meyer and Haerter, 2020). We note that the median value of T_{drop} is -2.9 K and that of Δu_{max} is $+4.4 \text{ m} \cdot \text{s}^{-1}$. Comparing a linear least-squares fit (green line), constrained to passing through zero ($\Delta u_{\text{max}} = 0$ should correspond to $T_{\text{drop}} = 0$), with a square-root least-squares fit (red curve, Equation 3) by inspecting the residuals (Figure 7b,c) indicates that the latter is a more appropriate fit, given that there is no trend in the residuals of the square-root fit.

By assuming a fixed internal Froude number k = 0.7, the fitting constant a = 478 m can be understood as an estimate of the CP height *H* (Equation 3). This estimate is highly sensitive to the chosen value for the Froude

number k. Underestimating k will lead to an overestimation of H (and vice versa). Nevertheless, the estimated value is comparable to the CP height estimate based on the temperature anomaly extrapolation $z_0 \sim 500$ m discussed earlier. Our value is larger than the average 300-m CP heights over tropical oceans inferred from aircraft pressure measurements (Terai and Wood, 2013) and lower than the height of 746 m found for CPs over Hamburg, Germany, with a pressure-anomaly based extrapolation from tower measurements (Kirsch *et al.*, 2021). We point out that our CP height value is much lower than the 1.5–2 km heights found for early simulated thunderstorm outflows (e.g., Droegemeier and Wilhelmson, 1987; Liu and Moncrieff, 1996).

3.3 | How does rain intensity influence CP strength?

CPs develop primarily through the evaporation of rain in the subcloud layer (Kurowski *et al.*, 2018). The evaporation is enhanced in rain showers with high drop number density, though it decreases with high ambient relative



FIGURE 7 Cold pool property relationship compared with the theoretical model. (a) Gust front strength Δu_{max} versus relative temperature drop T_{drop}/T_0 for all CPs detected during 2010–2019. Considering Equation 3, the fitting constant *a* in the square-root fit can be understood as an estimate of the CP height. In the fitting functions, x corresponds to the values on the x-axis. (b) Residuals from square-root fit (Equation 3), trend shown as a thin dotted black line. Approximately symmetric spread around zero (red line). (c) Same as (b), but for linear fit. Systematic tendency with an underestimation (overestimation) of Δu_{max} for low (high) T_{drop}/T_0 [Colour figure can be viewed at wileyonlinelibrary.com]

humidity (Seifert, 2008). We here test this bivariate relationship between the CP, the parent rain cell, and their mutual environment.

Relative humidity can be estimated directly from the measured dew-point depression $\tilde{T} \equiv \overline{T} - \overline{T}_d$, where \overline{T} and \overline{T}_d are averages of temperature and dew-point temperature over the 50-min time interval preceeding CP detection ([$t_0 - 60, t_0 - 10 \text{ min}$]) at the 140-m level of Cabauw tower. Whereas the data used in this study do not contain explicit microphysical information, rain intensity, *I*, is a rough proxy for rain drop number density. Its spatial average over the entire rain cell area, denoted $\langle I \rangle$, is computed at a single time step t_{RC} (Section 2.3).

While crude, this spatial average lowers the sensitivity to biases of the radar measurement, such as that due to the presence of ice, artificially increasing the reflectivity locally.

Using linear regression on the data from the ten summers of CP data (2010–2019), we expand the temperature drop T_{drop} in terms of the low-order terms:

$$T_{\rm drop} \sim \alpha_0 + \alpha_1 \langle I \rangle + \alpha_2 \tilde{T} + \alpha_3 \langle I \rangle^2 + \alpha_4 \tilde{T}^2 + \alpha_5 \langle I \rangle \tilde{T} + \mathcal{O}(3).$$
(4)

The regression analysis indicates that significant nonlinearity enters through the quadratic terms of rain intensity $\langle I \rangle^2$ and \tilde{T}^2 , whereas the mixed term $I_{\text{mean}}\tilde{T}$ shows nonsignificant correlation, that is, α_5 shows very high standard error, and is thus neglected. To avoid the simple model predicting CPs at vanishing rain intensities or in a totally saturated atmosphere, we impose the physically meaningful restriction of zero intercept, that is, $\alpha_0 = 0$. Together, we retain the fit function:

$$T_{\rm drop} \sim \alpha_1 \langle I \rangle + \alpha_2 \tilde{T} + \alpha_3 \langle I \rangle^2 + \alpha_4 \tilde{T}^2 + \mathcal{O}(3).$$
 (5)

We here compare a linear regression, where $\alpha_3 = \alpha_4 = 0$, and a nonlinear regression, where α_3 and α_4 may vary (Table 1). In both models, the CP strength is positively correlated with the dew-point depression \tilde{T} and mean rain intensity $\langle I \rangle$, confirming that CPs are measurably strengthened by drier environmental conditions and larger rain intensities. In the nonlinear regression, all four remaining regression coefficients are found to be statistically significant. We speculate that the negative dependence on $\langle I \rangle^2$ ($\alpha_3 < 0$) stems from very strong rain quickly saturating the subcloud atmosphere, thus diminishing further rain evaporation.

To give a more tangible interpretation of the linear multivariate regression, we note that, with the coefficients in Table 1, a CP becomes 1 K colder if the rain intensity of its parent cell is incremented by $4 \text{ mm} \cdot \text{hr}^{-1}$ (instantaneous area-averaged rain intensity) or if the ambient relative humidity is increased according to a 2.5-K larger dew-point depression. We compare our results with a recent study by Kirsch *et al.* (2021) for CPs over Hamburg, Germany, hence a similar geographic region. The study shows two separate regressions, to determine the relationship between CP temperature perturbation and point-measured accumulated rainfall and CP temperature

TABLE 1Fitted coefficients for the linear and nonlinearregression models (Equation 5)

	Linear regression		Nonlinear regression	
	Coeff.	Std. error	Coeff.	Std. error
$\alpha_1 [\text{K} \cdot \text{hr} \text{mm}^{-1}]$	0.25	0.03	0.44	0.09
<i>α</i> ₂ [-]	0.40	0.03	0.51	0.08
$\alpha_3 [\text{K} \cdot \text{hr}^2 \cdot \text{mm}^{-2}]$			-0.020	0.007
$\alpha_4 [\mathrm{K}^{-1}]$			-0.021	0.007
R-squared (uncentered)	0.89		0.92	

Note: α_1 and α_2 represent the coefficients of the terms linear in rain intensity and temperature, respectively, whereas α_3 and α_4 represent the coefficients of the terms quadratic in rain intensity and temperature, respectively.

perturbation and pre-event saturation deficit. CP strength is found to increase with increasing point-measured rainfall and with increasing pre-event saturation deficit, in line with our model.

4 | CONCLUSION

In this study, we design and validate a methodology to detect convectively generated cold pools (CPs) and their gust fronts over the Netherlands and relate them to their parent rain cell and environment. CP characteristics have been studied over many years from an observational point of view; however, there are very few statistical studies of CPs over midlatitude coastal land. Our study stands out by combining tower and radar measurements to analyze more than 100 CPs in relation to their generating rain cells. We wish to highlight the following findings:

- The patterns in the horizontal and vertical wind measurements confirm the existence of a vortex ring in the CP head.
- The detected CPs show weak or absent moisture rings, while the CP interior shows a negative moisture anomaly.
- A simple model consisting of a multivariate linear combination of pre-event dew-point depression and area-averaged rainfall intensity allows a prediction of the generated CP's strength.

To study the evolution of CPs in time and their properties in relation to their parent rain cell and environment, we use the combination of local measurements from a 213-m meteorological tower located in Cabauw, the Netherlands, and precipitation radar output. To identify CPs from time series of point measurements, an existing algorithm for detecting CPs over the ocean (de Szoeke et al., 2017) is extended to capture CPs over land by (a) an increased threshold on the temperature anomaly (1.5 K), and (b) an additional criterion on the horizontal wind-speed anomaly. Studying a few exemplary CPs revealed a vertically coherent signal in horizontal wind velocity ("wind peak") at the tower during the passage of a CP gust front. Therefore, in our algorithm we record events as CPs if they are characterized by both a temperature drop and a coherent wind peak. This allows us to isolate the CPs from the temperature fluctuations over land, which come without, or with a very weak, wind signal. The algorithm shows low sensitivity to the threshold on temperature anomaly but high sensitivity to the wind-peak threshold. We thus recommend keeping the threshold on temperature at 1.5 K, whereas the threshold on the wind peak should be varied in accordance with the threshold on precipitation intensity of the parent rain cells under consideration. Our choice of parameters may bias the algorithm to detecting only strong CPs, the fronts of which are often found to be very close to the parent rain cell at the time of detection, indicating a low age of the CPs at the time of detection or a squall-line-like system, where the CP is advected along with the cloud. However, we find that, by this constraint, it is ensured that all identified cases can be attributed to the passage of a CP rather than other forms of fluctuations. Our method's reliability is confirmed by the associated updrafts and succeeding dry moisture anomalies measured for all detected CPs.

The composites of 189 CPs from measurements taken during ten summers (May-September 2010-2019) allow us to study statistical CP properties and their radial structure. The CP gust front is characterized by a strong updraft, which coincides temporally with the beginning of the negative temperature anomaly and precedes the positive horizontal wind anomaly and negative moisture anomaly. The presence of a vortex ring on the edge of the CPs is confirmed by the changing signal in horizontal wind measurements at different heights, in the window of time around the CP passage. In contrast to studies of oceanic CPs (de Szoeke et al., 2017; Zuidema et al., 2017), we do not detect a clear signature of moisture rings, that is, areas of elevated moisture, on the edges of the CPs. This might be due to the smaller magnitude of latent heat fluxes over land with respect to the tropical ocean (Drager et al., 2020). The absence of CP moisture rings in our observations suggests that, for our ensemble of CPs, the thermodynamic triggering of new rain events is less important than the mechanical triggering of new rain events, particularly driven by the collision of strong gust fronts (Tompkins, 2001a; Torri et al., 2015; Drager et al., 2020). The interior

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of the CPs in our analysis is characterized by enhanced horizontal wind anomalies and longer-lasting cold, dry air. The dry anomaly stands in contrast to a recent study of CPs over Germany (Kirsch et al., 2021), which finds enhanced moisture in CP interiors. The thermodynamic anomaly recovers within two hours of the passage of the gust front, which is clearly shorter than the convectively active time in a day. It is, therefore, possible for multiple CPs to occur in the same location within one day, despite the inhibiting effect of a cold, dry lower boundary layer on the formation of new convection. We do indeed, in some cases, detect more than one CP on the same day. Regarding the scale of CPs, our current results indicate that typical dynamical gust fronts have a width of 1 km. For the CP height, we extrapolate two values from (i) the vertical gradient in temperature anomalies measured at different tower levels, and (ii) the relationship between CP temperature anomaly and CP speed. Both estimates (\sim 500 m) are lower than the typical CP height over continental land and higher than the typical CP height over the ocean. This might be due to the coastal location, which provides an interesting environment in which to study CPs that are neither fully oceanic nor fully continental. Furthermore, the examination of radar images and attribution of generating rain events reveals that some CPs that we detect emerge from large-scale weather patterns with lower rain intensity than convective thunderstorms. We suspect that this contributes weaker and shallower CPs to the ensemble.

By combining the tower measurements with the radar data and using a wind-direction criterion, we were able to attribute a parent rain cell to the majority of CPs detected. This allows us to confirm the positive relation between CP strength and precipitation intensity with a bivariate, linear regression of CP temperature anomalies against the pre-CP dew-point depression and rain intensity averaged over the full rain-cell area. Our simple model allows the prediction of CP strength (temperature anomaly), conditioned on both microphysical and environmental parameters. Knowledge of the relation between a given population of rain cells and the CPs initiated by any of them is important as a benchmark for numerical simulations, both idealized studies and comprehensive high-resolution climate models.

4.1 | Outlook

Numerous recent studies have highlighted the importance of CP effects in structuring the convective cloud and precipitation field over space and time (Rio *et al.*, 2009; Böing *et al.*, 2012; Schlemmer and Hohenegger, 2014; Böing, 2016; de Szoeke *et al.*, 2017; Haerter and Schlemmer, 2018; Haerter *et al.*, 2019; 2020). Further observational studies along the lines presented here may help clarify pressing modelling questions, such as how parameterized CPs are affected by large-scale weather conditions, or the numerical grid resolution required to resolve the gust fronts of spreading CPs appropriately. Future research is required to study the last step in the causal chain of convection, namely the attribution of the triggering of new convective cells to the detected CPs and their parent rain cells. The current study, and work following up on it, can help improve the realism of models and decipher how mesoscale convective systems build up dynamically and what role CPs play in correlating the dynamics of the individual rain cells involved.

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AUTHOR CONTRIBUTIONS

Irene L. Kruse: conceptualization; data curation; formal analysis; methodology; validation; visualization; writing – original draft; writing – review and editing. **Jan O. Haerter:** funding acquisition; investigation; project administration; supervision; writing – original draft; writing – review and editing. **Bettina Meyer:** conceptualization; methodology; supervision; validation; writing – original draft; writing – review and editing.

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APPENDIX A

We here present three additional figures. Figure A1 shows the relative distance between detected CP gust fronts and the rain cell edge for all detected CPs, to highlight that many of the CPs detected are close to their generating rain cell.

Figure A2 shows rain intensity of some of the generating rain events, from radar data, exemplifying the different kinds of rain events producing the CPs we are detecting.

Figure A3 shows the general time of occurrence of the detected CPs. Note that the peak is in the afternoon (14-17 local time, 12-15 UTC), following the diurnal convective cycle.



FIGURE A1 Relative distance *d* between detected CP gust front and the rain-cell edge for all CPs detected in the 10 summers 2010–2019. The CP gust front is detected directly at the edge of the rain cell for d = 1, ahead of the rain cell for d > 1, or below the rain cell, that is, it rains at the tower's location, for d < 1 [Colour figure can be viewed at wileyonlinelibrary.com]



FIGURE A2 Rain intensity from radar data, exemplifying different weather situations at moments of CP detection. Note the proximity of the rain cells to the Cabauw tower (red dot). Red contours indicate the intensity threshold used to track the rain cells. (a) Isolated rain cells (August 19, 2019, 15:25); (b) squall-line-like convective cell (May 8, 2019, 15:25; (c) extensive front approaching from the Atlantic (August 24, 2018, 21:30); (d) mesoscale convective system (June 7, 2019, 15:10). In cases (a) and (b), the isolated rain cells with well-defined COM allow a unique attribution, whereas in cases (c) and (d) the COM can be ill-defined and the attribution is more ambiguous [Colour figure can be viewed at wileyonlinelibrary.com]



FIGURE A3 Time of occurrence of all CPs detected in the 10 summers 2010–2019. Note that the local summer time in the Netherlands is CEST, corresponding to UTC+2 [Colour figure can be viewed at wileyonlinelibrary.com]