Observations of Kinematics and Thermodynamic Structure Surrounding a Convective Storm Cluster over a Low Mountain Range

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ABSTRACT

Measurements of a convective storm cluster in the northern Black Forest in southwest Germany have revealed the development of a warm and dry downdraft under its anvil cloud that had an inhibiting effect on the subsequent development of convection. These measurements were made on 12 July 2006 as part of the field campaign Prediction, Identification and Tracking of Convective Cells (PRINCE) during which a number of new measurement strategies were deployed. These included the collocation of a rotational Raman lidar and a Doppler lidar on the summit of the highest mountain in the region (1164 m MSL) as well as the deployment of teams carrying radiosondes to be released in the vicinity of convective storms. In addition, an aircraft equipped with sensors for meteorological variables and dropsondes was in operation and determined that the downdraft air was approximately 1.5 K warmer, 4 g kg\(^{-2}\) drier, and therefore 3 g m\(^{-3}\) less dense than the air at the same altitude in the storm’s surroundings. The Raman lidar detected undulating aerosol-rich layers in the preconvective environment and a gradual warming trend of the lower troposphere as the nearby storm system evolved. The Doppler lidar both detected a pattern of convergent radial winds under a developing convective updraft and an outflow emerging under the storm’s anvil cloud. The dryness of the downdraft air indicates that it had subsided from higher altitudes. Its low density reveals that its development was not caused by negative thermal buoyancy, but was rather due to the vertical mass flux balance accompanying the storm’s updrafts.

1. Introduction

The evolution of a convective storm strongly depends upon the interaction with its environment. Relevant influences of the atmospheric environment include the vertical distributions of temperature, moisture, wind velocity, and aerosols (e.g., Khain et al. 2005). Another constraint is the nature of the topography. Over low mountains secondary flows can develop during daytime (Koßmann et al. 1998) that are capable of triggering convective cells (Barthlott et al. 2006). Inadequate simulation of these processes leads to significant systematic errors in quantitative precipitation forecasting (Schwitalla et al. 2008). The interaction also includes the influences of the convection on its own atmospheric surroundings. For example, the evaporation of hydrometeors produced by the storm results in cooling of subcloud air (Byers and Braham 1948). Additionally, low-level cooling often occurs across a larger area as a result of shading by the storm’s anvil cloud (Markowski et al. 1998; Marsham et al. 2007). Other processes include the interaction of convective up- and downdrafts with the flow in the mid- and upper troposphere, and with stable, dry layers or lids (Morcrette et al. 2007). A major part of this case study concerns yet another interaction, namely, that which resulted in a dry and warm downdraft that has positive thermal buoyancy. It formed as a response to the upward mass flux associated with the convective updrafts. Observations of such compensatory downdrafts occurring in the mid- and upper troposphere have previously been presented and discussed by Sinclair (1973), Fritsch (1975), and Yuter and Houze (1995). We will present here measurements...
of the effects of such downdrafts on the lower troposphere.

Recognizing the importance of understanding the interaction between a storm and its environment for quantitative precipitation forecasting, many field experiments have been carried out worldwide to collect observational data. Recent field campaigns are, for example, the International H2O Project (IHOP_2002), which took place over the southern Great Plains of the United States in 2002 (Weckwerth et al. 2004), the Convective Storm Initiation Project in the southern United Kingdom in 2005 (Browning et al. 2007), and most recently the Convective and Orographically-induced Precipitation Study (COPS) in southwestern Germany and northeastern France (Wulfmeyer et al. 2008). A major challenge of such measurement campaigns is to collect atmospheric data at exactly the right time and location to capture the processes under study.

The data presented in this paper have been collected in the framework of the measurement campaign Prediction, Identification and Tracking of Convective Cells (PRINCE) that was carried out in the northern Black Forest of southwest Germany in July 2006. Here we report on the observations of 12 July 2006 and their interpretation. Trentmann et al. (2008) report on efforts to simulate the convective evolution on that day using three state-of-the-art convection-allowing models. The goals of PRINCE were to investigate the following: 1) why a storm system would develop at the time and location where it was observed; 2) in what ways its environment influenced its development; and 3) in what ways the storm itself influenced its environment and which effects these had on the subsequent convective evolution. In section 4 we will demonstrate to what extent the collected data allowed us to answer these questions.

During PRINCE, a number of new measurement strategies were deployed. First, two lidars were positioned on the summit of a mountain. A scanning Doppler lidar measured the wind field, while a scanning rotational Raman lidar measured atmospheric temperature, the particle backscatter coefficient, and the extinction coefficient at a wavelength of 355 nm. Second, a research aircraft was equipped with instruments in order to receive real-time data of meteorological satellites and data from ground-based operational weather radars. This allowed the aircraft to modify the flight track so that data in the vicinity of developing convective cells could be sampled. Finally, mobile teams equipped with radiosondes were deployed in order to release weather balloons in the spatial and temporal vicinity of the most interesting convective processes.

2. Measurement systems

The PRINCE measurement campaign was carried out in the northern Black Forest in southwestern Germany, a low mountain range flanked by the Rhine Valley to its west and an elevated hilly landscape on its east
(Fig. 1). Its highest peak, Hornisgrinde, is found in its western part and has its summit at 1164 m above mean sea level (MSL). A number of winding valleys extend from the Rhine Valley (elevation of 100–200 m MSL) into the mountain range, including those through which the Murg and Rench Rivers flow. In these valleys, pronounced valley and mountain breezes occur on days with strong insolation (Barthlott et al. 2006) as was the case on 12 July 2006, the day discussed in this paper. Various sensors were placed on the summit of Hornisgrinde. The individual ground-based and airborne sensors that were operated are described in sections 2a–e.

a. **Rotational Raman lidar**

A full description of the rotational Raman lidar of the University of Hohenheim is given by Radlach et al. (2008). During PRINCE, the lidar transmitter consisted of a Nd:YAG laser that emits pulses at the frequency-tripled radiation of 355 nm with a pulse energy of about 300 mJ and a repetition rate of 30 Hz. A scanner points the laser beam in any direction of interest. Radiation backscattered by molecules and particles is directed via the same scanner mirrors toward a 40-cm-diameter telescope. The elastically scattered signal and two signals out of the rotational Raman bands are separated with a sequential setup of three channels (Behrendt and Reichardt 2000). The signals are detected with photomultiplier tubes and digitized simultaneously in analog and photon-counting mode with a temporal and vertical resolution of 10 s and 3.75 m, respectively. From the ratio of the two rotational Raman signals the temperature can be calculated (Behrendt 2005). Also the particle backscatter coefficient and extinction coefficient can be determined independently with the elastic and rotational Raman signals (Behrendt et al. 2002).

b. **Doppler lidar**

A 2-μm Doppler lidar “WindTracer” was operated by Forschungszentrum Karlsruhe (FZK). It is a commercially available system produced by Lockheed Martin Coherent Technologies. The system sends out laser pulses with a 500-Hz repetition rate through a 2-axes scanner covering the entire upper hemisphere. Line of sight wind, signal-to-noise ratio, and aerosol backscatter are calculated by the system’s real-time data processing unit for 100 range gates selectable between 72- and 96-m lengths. Averaging 50 laser pulses yields a data rate of 10 Hz, covering at least the boundary layer up to a distance of 6–10 km depending on the aerosol load present. During PRINCE, a scan pattern consisting of the range–height indicator (RHI; i.e., varying elevation at a fixed azimuth) and the plan position indicator (PPI; i.e., varying azimuth at a fixed elevation) scans have been performed. The system had a high scan speed (6° s⁻¹) and repeated its scan pattern every 5 min in an unattended continuous operation mode.

c. **C-band Doppler radar**

The C-band Doppler radar Gematronik Meteor 360 AC is located at FZK, roughly 60 km north of Hornisgrinde. During PRINCE, it continuously repeated a 14-elevation-volume scan every 5 min, recording reflectivity and radial velocity data with a range resolution of 300 m and an azimuthal resolution of 1°.

d. **DO-128 research aircraft**

The two engine research aircraft Dornier 128–6, D-IBUF, is operated by the University of Braunschweig and FZK. It has a nose boom designed for meteorological measurements in undisturbed air (Corsmeier et al. 2001), which contains redundant sensors for the measurement of wind, temperature, and humidity. Using combined inertial navigation system (INS) and global positioning system (GPS) navigation, a sample frequency of 100 Hz is realized. With an approximate airspeed of 65 m s⁻¹, the spatial resolution of the measurements is less than 1 m. The accuracy of the horizontal wind speed is 0.5 m s⁻¹ for the u and v component and 0.1 m s⁻¹ for the vertical wind. For temperature, the error is 0.2 K, and the mixing ratio has a relative error of 4.8%. The time required for the temperature sensor to indicate 63.5% of a sudden temperature change is 0.5 s. Precise high-frequency humidity measurements are accomplished by complementary filtering of Lyman-alpha and Humicap data, thus reaching a time constant of 0.04 s. In addition to the onboard sensors, the aircraft is equipped to release up to 30 autonomous operating dropsondes (see section 2e) within a time frame of a few minutes.

e. **Radiosondes and dropsondes**

A radiosonde site was established on the western slope of Hornisgrinde at Brandmatt, Germany, approximately 3 km west of the peak of the mountain. Several Graw DFM97 sondes were released from this site at 3- and 2-h intervals during the daytime of 11 and 12 July, respectively. Additionally, radiosondes of the type developed by Kottmeier et al. (2001) were released by three mobile teams, and dropped from the DO-128 research aircraft. The mobile teams drove to preselected sites that had been identified as suitable for the release of radiosondes. On each day of the campaign, the PRINCE operations center decided to which of the sites the teams were sent, dependent on the expected weather development. On 12 July, stations 9, 17,
and 23 were selected. Their locations are shown in Fig. 1. A special characteristic of the sondes is that the measured data is not transmitted via a radio signal, but is stored internally on a flash memory device to be read out after recovery of the sondes. In order for the retrieval of the sondes after use, they are equipped with a mobile phone that sends its geographical coordinates, detected by a GPS receiver, as a short message service (SMS) message.

3. Observations

a. Synoptic setting

For a discussion of the synoptic-scale weather, analyses of the European Centre for Medium-Range Weather Forecasts (ECMWF) Integrated Forecast System (IFS) of 1200 UTC 11 July 2006 and 0000 and 1200 UTC 12 July 2006 are presented in Fig. 2. The potential vorticity (PV) maps show an upper-tropospheric trough, characterized by high PV that moved eastward over southern Scandinavia late on 11 July. At 0000 UTC 12 July 2006, the southern part of the trough had developed into a thin streak of high PV stretching southwestward from southern Sweden to Belgium. As the northern part of the trough moved on into northeastern Europe, the streak split and a local maximum of upper-level PV cut off over northern France, approaching the PRINCE area at 1200 UTC 12 July.

At 1200 UTC 11 July (Fig. 2a), the surface pressure field shows a ridge extending from the northern Bay of Biscay toward Denmark. Behind the passage of the trough, this ridge elongated a bit farther toward the Baltic Sea at 0000 UTC (Fig. 2c), before weakening and leaving the PRINCE area within a region of weak surface pressure gradients. In the temperature field, a zone of enhanced gradient can be seen to stretch from southern Scandinavia over north and west Germany toward the Bay of Biscay. At all times, the PRINCE area was on the warm side of this zone.

On the evening of 11 July 2006, isolated thunderstorms formed over the Vosges Mountains, the low mountain range to the west of the Rhine Valley. In the early morning of 12 July 2006, storm systems formed over the Rhine Valley about 30 km to the west-northwest of Hornisgrinde at around 0330 UTC. In what follows, we will focus on the development of storms during the daytime of 12 July, when storms formed over the Black Forest.

The fact that storms did not develop over the Black Forest on 11 July 2006, but did develop the following day, can be understood when considering the temperature and moisture profiles measured with sondes released at Brandmatt on both days (Fig. 3). At 1303 UTC 11 July 2006, a warm layer around 775 hPa acted as a strong lid preventing thermals from the boundary layer to penetrate into the free atmosphere. Convective inhibition (CIN) at that moment was 88 J kg$^{-1}$ for a 50 hPa mixed-layer parcel (Table 1). At 1900 UTC, the air at 775 hPa and above was 1°C–2°C colder so that CIN was reduced despite the formation of a shallow stable surface layer. This is likely a result of upward vertical motion induced by the approach of the upper-level vorticity maximum. At 0702 UTC 12 July 2006, despite nocturnal cooling having reduced near-surface temperatures, CIN still had approximately the same magnitude. It would have been much higher if the air between 850 and 550 hPa had not again cooled significantly. Most CIN appears in the lowest 2 km AGL and can be expected to disappear by diurnal heating. Unlike the profiles of 11 July, no well-defined warm layer that could inhibit deep convective development was present above 2 km AGL, except for an inversion at 530 hPa (about 5.5 km MSL). This inversion, which was evident in later ascents from Brandmatt and in the 1200 UTC radiosonde data from nearby Nancy and Stuttgart (not shown here), would later indeed influence the development of convective storms.

b. Preconvective conditions

Shortly after 0800 UTC 12 July, a well-mixed planetary boundary layer (PBL) was present over the summit of Hornisgrinde. Figure 4 shows the particle backscatter coefficient as measured with the rotational Raman lidar during the time interval from 0813 to 0959 UTC.

In the lowest 800 m above ground, aerosol-rich thermals can be identified as subtle, vertically coherent structures of relatively high backscatter coefficient. Both higher relative humidity and a higher aerosol content within the thermals are possible causes for this. An undulating transition, labeled A, is indicated by a line of small squares. There the general increase of particle backscatter with altitude stops abruptly. It separates the turbulent air below from smoother structures above, and is hence thought to represent the top of the boundary layer. The increase of backscatter with height in the boundary layer is likely due to the increase of relative humidity with height, which can be expected in a well-mixed air mass. The higher relative humidity results in hygroscopic growth of aerosols and thus a higher backscatter coefficient. This transition A oscillates twice, before a first shallow convective cloud is detected directly over the lidar at 0835 UTC. We are not able to infer from the data what triggered these oscillations.

Above the boundary layer, the particle backscatter coefficient is weaker. A comparison with the sounding released at Brandmatt at 0900 UTC shows that this air is drier. A shallow aerosol-rich layer (labeled B) oscillates in phase with the underlying boundary layer.
around the level of 1100 m above the summit. Farther up, a moist zone stretching up to about 1400 m above the summit is followed by a 200-m-thick layer of weak backscattering signal (labeled C) that was relatively dry and warm. Finally, within the middle of layer in D, a weakening of the backscatter coefficient with height is found, which coincides with a drop in relative humidity measured by the radiosonde.

FIG. 2. Analyses of the operational ECMWF IFS. (left) Mean sea level pressure (hPa) and temperature (°C) at the 850-hPa level and (right) potential vorticity (grayscale, in potential vorticity units, i.e., 10^{-6} m^2 s^{-1} K kg^{-1}) at the 330-K level at (a), (b) 1200 UTC 11 Jul 2006; (c), (d) 0000 UTC 12 Jul 2006; and (e), (f) 1200 UTC 12 Jul 2006. A small square near the center of each map indicates the PRINCE area (Fig. 1).
In the 1.5 h after the first convective cloud was detected by the lidar, the various layers of high and low aerosol backscatter start to become less distinct: their smooth appearance becomes more convoluted. Our interpretation is that the deep convection that initiated to the east of the mountain during that period, influenced the flow and the observed aerosol distribution above the summit. Lower clouds that were associated with the developing storm system, started to appear at 0948 UTC. The Doppler lidar was located next to the rotational Raman lidar. A scan from west (left) through the zenith to the east (right) at 0933 UTC is shown in Fig. 5. It displays several layers with different velocities. To the east of the summit, winds with an easterly component of about 1–2 m s\(^{-1}\) are observed throughout most of the layer in which data was available. On the west (left) side of the summit, the flow pattern is quite different: a bottom layer of winds with a 1 m s\(^{-1}\) westerly component is found, which has a thickness of approximately 700 m. It represents the upslope branch of the mountain-breeze system. At some distance from Hornisgrinde, this layer disappears below the lowest possible elevation (0\(^{\circ}\)) of the lidar. A 500–700-m-thick layer above this surface layer has an easterly wind component of 2–3 m s\(^{-1}\). Except for a weak flow to the north, no near-surface flow away from the mountain top can be detected in velocity–azimuth display (VAD) scans (i.e., a radial velocity scan with varying azimuth at fixed elevation) of the Doppler lidar (not shown). This supports the idea that the air within the upslope flow out of the Rhine Valley rises near the mountain top, and returns toward the valley as part of this elevated easterly flow. The overall easterly flow, which is also detected east of Hornisgrinde, likely contributes to it as well. A comparison of the maximum altitude of this layer (about 1000 m AGL) with that of transition A in Fig. 4, shows that the

![Skew T–log\(p\) diagram showing radiosonde ascents from Brandmatt (48.62°N, 8.16°E) at 1303 UTC 11 Jul 2006 (red), 1900 UTC 11 Jul 2006 (blue), and at 0702 UTC 12 July 2006 (green). The dotted line in each color represents the dewpoint temperature, whereas the continuous line is the temperature. The dashed colored lines are the temperature curves of lifted parcels originating from a 50-hPa thick mixed layer right above the surface in Brandmatt. On the right-hand side an approximate height scale is shown, as well as the horizontal wind direction and velocity (half barb = 2.5 m s\(^{-1}\); whole barb = 5 m s\(^{-1}\)).](image)

<table>
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<tr>
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<tr>
<td>0702 UTC 12 Jul</td>
<td>62</td>
<td>994</td>
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top of the elevated easterly flow is at approximately the same altitude as the top of the boundary layer.

c. Convective initiation

At around 0830 UTC, the first convective clouds were visually observed across the western and eastern slopes of the Murg Valley. At 0924 UTC, the C-band radar detects the first hydrometeors produced by deep convective clouds across the hills east of the Murg Valley (not shown). The evolution of the radar echoes on 12 July 2008 are presented in Fig. 6. At 1000 UTC, weak echoes of hydrometeors are also detected to the north-
FIG. 6. Reflectivity (dBZ) measured by the C-band radar located at the FZK (in Fig. 1) in 30-min intervals. The images display the (bottom left) maximum reflectivity observed in a vertical column projected on the $x$-$y$-plane map showing the topography, (top) the maximum observed in a north–south slice projected on the $x$–$z$ plane, and (right) the maximum measured in a east–west slice projected on the $y$–$z$ plane. The $z$ coordinate is given in kilometers MSL. The narrow black lines are the projections of the flight pattern of the DO-128 research aircraft on the respective planes. The big black rectangle denotes the PRINCE area in Fig. 1. The red dot indicates the summit of Hornisgrinde and the green dot indicates the radiosonde station number 17.
The situation at 1010 UTC is depicted in Fig. 7. At that moment, the convective storm that initiated east of the Murg Valley has broken through the inversion at about 5-km altitude, for reflectivity of over 40 dBZ extends up to 10 km AGL. However, the system west of the Murg Valley does not grow as quickly. Only weak reflectivity is detected above 5 km MSL. Photographs taken from the city of Karlsruhe (not shown) suggest that the eastern system rapidly broke through the inversion whereas the western one was significantly hindered by it: its top spread out horizontally in the midtroposphere.

At 1008 UTC, a new small cell, marked with an arrow in Fig. 7, developed just to the northeast of Hornisgrinde. This cell merged with the existing western convective system during the following half hour. The fact that the cell developed close to the Doppler lidars at Hornisgrinde enables us to study the associated low-level wind field. Figure 8 shows the radial wind component, measured in a scan at 4° elevation. Two boundaries of abrupt changes of radial velocity in the radial direction are indicated, labeled A and B. Along boundary A, inward velocities close to the radar change to outward velocities farther away from the lidar. This is easily explained by noting that the lidar, when pointing westward, samples the upslope branch of the valley breeze system close to the radar and the return branch at greater distances.

To the east of the lidar, the wind field has changed considerably in comparison with that measured at 0933 UTC (Fig. 5). Directly to the east of the lidar, winds...
now have a westerly component. Along boundary B, located at about 5 km from Hornisgrinde, an outbound-to-inbound direction change occurs implying that radial velocities converge along it. Although tangential velocity components are not known, the collocation with the reflectivity core above the same location is striking, and strongly suggests that low-level mass convergence was taking place.

d. The mature and decaying storm system

Starting at 1033 UTC, the Doppler lidar scans indicate a shallow outflow from the storm system crossing Hornisgrinde. An RHI scan along the east–west plane at 1103 UTC shows the flow from east to west across Hornisgrinde (Fig. 9) at lower altitudes (up to ~500 m AGL). A clear westerly velocity component toward the storm was observed above that layer. The interface between the easterly flow away from the storm system and westerly flow toward it slopes down to the west.

The sonde released at Brandmatt at 1106 UTC (Fig. 10, red) shows weak northeasterly winds between 900 and 850 hPa, consistent with the weak outflow of the storm system observed by the Doppler lidar. This layer has a higher potential temperature than the air present below 900 hPa in the Rhine Valley, which still has a weak component toward the storm. As a result, the outflow spreads out on top of this layer. The temperature minimum observed near 700 hPa is thought not to represent the actual air temperature, but rather the result of cooling of the temperature sensor owing to the evaporation of cloud droplets that had accumulated on the sensor while crossing a cloud layer near 730 hPa.

Initially, the outflow air between 900 and 850 hPa was rather moist. The 1106 UTC ascent indicates an average specific humidity near 11 g kg$^{-1}$ in that layer. At 1259 UTC, however, the outflow contained much less moisture, on average about 9 g kg$^{-1}$. The sounding released at 1502 UTC indicates that even hours after the storm system had dissipated, the low-level moisture did not recover to the level of 1106 UTC.

When the development of the storm system was well under way, the DO-128 research aircraft released six dropsondes from an altitude of about 6300 m MSL. The sondes sampled the environment below the anvil cloud of the storm and a few developing convective updrafts. The location of four of the sondes, numbered 1–4, are
plotted in Figs. 1 and 6d. The profiles of equivalent potential temperature $\theta_e$ and the saturated equivalent potential temperature $\theta_{es}$, have been displayed in Fig. 11. The variable $\theta_e$ has been selected because of its property to be approximately conserved under both dry and moist adiabatic vertical motions. The $\theta_{es}$ has been plotted in order to find out where a sonde penetrated a cloud. Moreover it is a measure of temperature. From the definition of $\theta_{es}$, it follows that the air is saturated where both curves overlap, which corresponds to a good approximation with areas containing clouds. We discuss below the profiles of the various sondes and therein refer to the features A–H indicated in Fig. 11.

Sonde 1, that was released just to the east of the storm system, shows a layer of warm, unsaturated air (labeled A) above 5600 m. This layer is situated upon a layer of slightly cooler, almost saturated air (labeled B). We can compare the profile with that of the preconvective environment, as sampled by the radiosonde released at Brandmatt at 0900 UTC, which is displayed on the right in Fig. 11. It can be seen that the air at A is warmer, having a $\theta_{es}$ of 68°C, than the air sampled ear-
lier at Brandmatt at the same altitude, which had a $\theta_{es}$ of 58°C. The high temperature of the air sampled by dropsonde 1 can be explained by the adiabatic warming of air that descended below the storm’s anvil cloud as part of compensating downdrafts of the thunderstorm. The lack of drying that one would expect to see in subsiding air, can be explained by the evaporation of hydrometeors falling out of the anvil cloud into this layer. The saturated layer in B between 5600 and 4800 m likely originates from the spreading out of a part of the convective updraft against the inversion that was present in the preconvective environment. In the Brandmatt sonde, this inversion is visible at 5200 m. A cloud deck like B is visible in the bottom right of Fig. 12, which shows a photograph taken from the research aircraft while flying at about 6300 m AGL, a few minutes before sonde 1 was released.

Below 4800 m, the profile of sonde 1 has more or less the same properties as the undisturbed environment, with the exception of the feature labeled as C. There, the sonde briefly traveled through a large cumulus cloud containing air originating from the boundary layer. Just below the maximum of $\theta_e$ and $\theta_{es}$ the former briefly exceeds the latter, which suggests that the air was supersaturated. That, however, has not necessarily been the case, as it was likely an undesired artifact of the instrument caused by the accumulation of drops of liquid water on the dewpoint mirror.

Below the cloud element, the sonde again falls through air having temperatures not too different from those of the preconvective environment (D), except that the air is slightly moister. In the lowest 200 m above the surface, a shallow layer of somewhat cooler, drier air (E) is observed. The wind within this layer was from north to northeasterly directions and had no component away from the storm, so that the outflow of the storm is an unlikely cause of its dry and cool properties. We hypothesize that the lower temperatures of the layer (E) are a result of radiational cooling, caused by shading by the anvil cloud, that, at the time of the measurement, had already blocked direct sunshine for about 2 h.

The profile measured by sonde 2, which was released about 3 km farther westward, has lower temperatures in region A and does not feature regions B and C. Like sonde 1, this sonde also detects cool, dry air near the surface.

Dropsonde 3 has been released within precipitation on the southeast side of the storm system. The profile displays a high relative humidity close to saturation be-
between 6500 and 3500 m. Below approximately 2300 m, a convective cloud element has likely been sampled. Below the saturated layer, which stretches down to 1800 m, a 700-m-thick layer of cool air is observed (F), with much lower values of $u_{es}$ than the previous two sondes indicated. Cooling by evaporation of hydrometeors is the most likely source of this and we believe that the cold pool of the storm system has been sampled here.

Dropsonde 4 has been released above a developing towering cumulus cloud. The sonde approached the cloud (G) at 4800 m, but was not able to penetrate it because of the large upward velocities associated with it. At some point the GPS device had detected an upward motion of 5 m s$^{-1}$. Considering that the typical fall speed of the dropsonde lies between 3 and 5 m s$^{-1}$, this implies that the air around the sonde was moving upward at 8–10 m s$^{-1}$. Indeed the difference of 10.0°C in $u_{es}$ between the cloud and the air outside of it corresponds with an actual temperature difference of 4.5°C, indicating that the air within the cloud had a high thermal buoyancy. Below 2800 m, the sonde again encountered a convective updraft (H) that had approximately the same $u_{es}$ as the convective cloud element G encountered before. Sonde 4, like sonde 3, samples cold and moist air associated with the storm system’s cold pool below 1600 m.

After releasing the dropsondes, the DO-128 research aircraft descended to 1150 m MSL and started to fly the pattern indicated by dashed lines in Fig. 1, starting in the northeast at 1139 UTC and reaching the end of the pattern at 1212 UTC. While doing so, temperature, moisture, and three-dimensional wind data—corrected for the motion of the airplane—were recorded. The respective temperature, moisture, and density data are shown in Figs. 13a–c. During this part of the flight, the storm system moved southeastward just quickly enough so that it moved out of the planned flight pattern before the aircraft arrived at its most southeastern location. The approximate location of the storm at 1200 UTC is indicated in Figs. 13a–c. The pattern of potential temperature shows a relative maximum with temperatures between 30°C and 31°C across the southwestern part of the probed area, whereas values are around 29°C farther to the north and east. The moisture shows a similar pattern with the highest moisture being observed where temperatures are lowest. In the northern and eastern parts of the pattern, the specific humidity was generally between 12 and 13 g kg$^{-1}$, while it drops to below 8 g kg$^{-1}$ in the central and southwestern parts. These data have been used to calculate the density of the air that was sampled using the equation of state, formulated as $\rho = \rho R_d T_v$, where $\rho$ is pressure, $\rho$ is density, and $R_d$ is the gas constant for dry air. Here, $T_v$ is the virtual temperature, which may be approximated as $T_v \approx T(1 + 0.608r)$, with $r$ representing the water vapor mixing ratio.

As the aircraft was at that moment flying through
considerably subsaturated air that did not contain any hydrometeors, the effects of liquid water need not to be included in the equation. The absence of hydrometeors has been confirmed by the aircraft crew and is consistent with the measurements of the C-Band radar. The resulting density data shown in Fig. 13c indicate that the air in the southwest was the least dense. The density difference between the southwest portion of the pattern and the north and east is about 3 g m$^{-3}$. Such a difference can be caused by a perturbation of 0.6 K in the temperature, 1.0 g kg$^{-1}$ in mixing ratio, or 2.5 hPa in pressure, which is much more than the measurement uncertainties of the sensors, which are 0.2 K, 0.5 g kg$^{-1}$ (at 10 g kg$^{-1}$), and 0.1 hPa, respectively. Hence, the observed low density cannot be attributed to measurement errors.

The observed winds show a divergent pattern (labeled D in Fig. 13) and confluence (labeled C in Fig. 13) to the east of the town of Forbach (Forb). Winds in the northern part of the flight pattern have an easterly component, while those to the east have a northerly component. This suggests that the northeasterly flow in the environment of the storm splits into two branches where it is blocked by the diverging dry and warm air over the Murg Valley. The dry and warm air must have descended from higher altitudes as horizontal advection of air within a relatively homogeneous air mass cannot result in local extrema of temperature or moisture. Moreover, the wind field depicted in Fig. 13 does not indicate that the dry air was advected toward the storm system. On the contrary, moist air is advected southwestward toward the storm, until it is blocked by the divergent dry air.

Another indication of downward motion taking place can be obtained from the radiosondes released at station 17. Data of two of the radiosondes, released at 1244 (red) and 1317 (blue) UTC are shown in Fig. 14. At both times the radiosonde station was located under the central part of the storm’s anvil cloud. The profile at 1244 UTC shows the existence of two warm layers labeled A and B, the latter being very dry (labeled C). At 1317 UTC, both layers have descended about 20 hPa or 200 m. Weak horizontal winds suggest that this is true local subsidence and not caused by horizontal advective effects. The vertical velocity that can be calculated by dividing the displacement of the layers by the elapsed time, is around 0.4 m s$^{-1}$, this estimate being valid at the altitude of those layers (i.e., around 2–4 km above the surface). In the absence of horizontal winds, the vertical wind component must vanish at the surface, so that it can be understood that the DO-128 research aircraft was not able to detect coherent downward motion when flying between 100 and 800 m AGL, its measurement error for vertical velocities being 0.1 m s$^{-1}$. Temperature

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**FIG. 13.** Measurements of (a) potential temperature, (b) specific humidity, and (c) air density measured by the DO-128 research aircraft between 1139 and 1212 UTC at an altitude of 1150 m MSL. The horizontal wind vector along the flight path is indicated with black arrows in (a)–(c). Hornisgrinde (H) is indicated by a red circle and radiosonde station 9 by a blue circle. The approximate location of the storm system at 1200 UTC (i.e., to the southeast of the flight pattern) is indicated. The areas labeled C and D indicate where confluence and divergence, respectively, of the horizontal wind field were observed.
measurements carried out by the rotational Raman lidar (Fig. 15) confirm that a similar lower-tropospheric warming occurred at Hornisgrinde. A comparison of the profiles of 0635, 0930, and 1435 UTC reveals a warming of the air above 2.7 km AGL. For example, at 3.8 km the temperature increases by 5 K in this period, most of it after 0930 UTC. Between 0930 and 1435 UTC some warming also occurs down to 1.8 km AGL. Between 1435 and 1710 UTC, after the decay of the convective system, the air above 2.7 km cools again by approximately 2 K.

An idealized cross section is sketched in Fig. 16. It displays the convective updrafts that generally occurred on the south (left) side of the system as it propagated to the south-southeast. Given the dryness of the midtroposphere, it is likely that considerable amounts of precipitation from the storm evaporated before reaching the surface, which has likely lead to cooling and localized downdrafts in vicinity of the strongest radar echoes. The warm, dry lower troposphere as observed by DO-128 between 1139 and 1212 UTC over the Murg Valley region and the warming observed between 1141 and 1241 UTC at station 9 is depicted as a large mesoscale downdraft that occurred under the anvil cloud of the storm between and to the north of the convective updrafts.

An important characteristic of the downdraft was that it did not reach down to the earth’s surface except at the top of Hornisgrinde and possibly some other hills and mountains. Indeed, warm and dry downdrafts are not commonly observed at the surface; unless they have very high downward momentum, as in so-called heat bursts, they do not reach the earth’s surface.

FIG. 14. Temperature, moisture, and wind profile depicted as in Fig. 3, as measured by radiosondes released at station 9 at 1141 (red) and 1245 UTC (blue).

FIG. 15. Temperature measurements with the rotational Raman lidar: profiles measured on 12 Jul 2006 at Hornisgrinde (1161 m MSL). The lidar data were acquired around the indicated time during (profile 1) 40, (profile 2) 30, (profile 3) 40, and (profile 4) 60 min, respectively. The spatial resolution is 37.5 m (profiles 1, 3, and 4) and 75 m (profile 2). The lidar signals are smoothed with the indicated window lengths. The data error increases with altitude from approximately 0.2 K at 1 km to 1 K at 5 km AGL.
e. **Subsequent convective development**

It turned out that no storms developed within the PRINCE area after the initial storm cluster dissipated, whereas they did at surrounding locations. Figure 17 shows two visible satellite images at 1200 and 1500 UTC. It can be seen that at 1200 UTC, the PRINCE cluster is one of the first storm clusters to develop, together with larger clusters in the southeast of the picture that develop over the Swabian Jura. Three hours later, the storm cluster has dissipated leaving some remnant cirrus fields over the Black Forest. At that moment, many new storms are initiating to the south, southwest, west, north, and northeast of the dissipating cluster. One may be tempted to attribute the suppression of convective initiation near the PRINCE cluster to the presence of a cold pool under the dissipating thunderstorm caused by evaporation of hydrometeors or reduced insolation. However, the soundings at Brandmatt (Fig. 10) and the rotational Raman lidar (Fig. 15) do not indicate any cooling of the boundary layer between 1200 and 1500 UTC. Insolation through the thin cirrus after the system’s decay has apparently been sufficiently strong to compensate any prior cooling, which leads us to conclude that shading by cirrus clouds has not been a dominant effect near Hornisgrinde. The sounding released at Brandmatt at 1502 UTC indicates that the moisture content of the boundary layer did not recover from the moisture drop that resulted from the downward flux of dry air. Additionally, the temperatures of the air directly above the boundary layer remained higher than before the convective cycle started. Both effects strongly reduced the buoyancy of a surface parcel in the first few kilometers above the boundary layer, compared with the preconvective situation. The satellite picture at 1500 UTC shows that convective development did occur at greater distances from the original storm cluster that were unaffected by the effects of its warm, dry downdraft.

4. **Discussion and conclusions**

During PRINCE, the coordinated deployment of synergistic remote sensing systems with in situ measurements was effective in revealing processes taking place during the life cycle of a convective storm cluster. The question of why convection ensued where it did can be answered only partially. Data from the ECMWF IFS and radiosondes suggest that the approach of an upper-tropospheric potential vorticity anomaly caused upward vertical motion that cooled the warm air capping the boundary layer the day before. The Doppler lidar has shown that a weak, but well-developed local circulation existed over the western slope of Hornisgrinde. High-resolution aerosol and cloud data by the Raman lidar in the preconvective environment revealed the existence of lower-tropospheric undulations, whose exact influence on convective initiation remains unclear at this time. Mass convergence was observed to the east of the summit in association with initiation of a convective updraft in which hydrometeors developed that were detected by the C-band radar. Although the occurrence of convective initiation cannot be fully explained, the present study underlines the value of state-of-the-art lidar systems for atmospheric studies.
An example of the influence the environment has had on the evolution is the interaction with the stable layer of air at around 5-km altitude, which constrained the depth of the western storm for a short time. More importantly, however, the study has shown how a storm changed its environment. The analysis of the various measurements indicates that a warm downdraft, with a diameter on the order of 10 km, formed under the decaying parts of the storm system. Its relative dryness, high temperature, and positive thermal buoyancy imply that cooling by evaporation of hydrometeors can be ruled out as the major forcing of the downward motion. Instead, it has primarily been a manifestation of the compensation of the upward mass flux in nearby convective updrafts. Further insight into the processes leading to convection initiation and organization may be provided by the assimilation of the PRINCE datasets, as demonstrated for other campaigns in Wulfmeyer et al. (2006) and Grzeschik et al. (2008).

Although most convective parameterization schemes take the warming and drying influence of such downdrafts into account (e.g., Kain and Fritsch 1990), observations of warming and drying in the lower troposphere have to our knowledge not yet been documented. Observations of warm downdrafts in the vicinity of convective storms at higher altitudes are presented by Fritsch (1975), who discusses observations by Sinclair (1973) made in the vicinity of severe storms at altitudes between 9 and 10 km AGL. The downward velocity within those were up to 5 m s$^{-1}$, and the entire downdraft had a horizontal extent of 13-km distance from the edge of the storm cloud. Yuter and Houze (1995) in a study of (sub)tropical thunderstorms across Florida, make a distinction between upper- and lower-level downdrafts, which have distinct dynamical origins. The upper-level downdrafts are caused by pressure gradient forces required to maintain mass continuity in the vicinity of updrafts, and lower-level downdrafts are caused by evaporation of precipitation. This paper complements these observations by demonstrating that such downdrafts may extend into the lower troposphere above the boundary layer, where they spread out and can cause significant inhibition to new convective initiation.

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