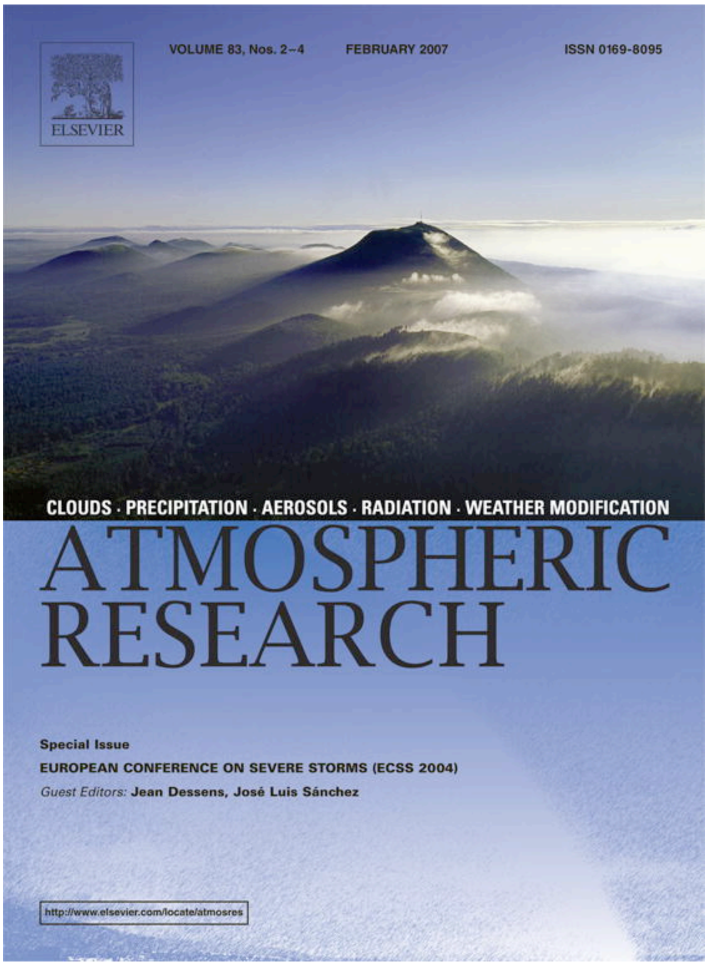
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Sounding-derived parameters associated with large hail and tornadoes in the Netherlands

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Abstract

A study is presented focusing on the potential value of parameters derived from radiosonde data or data from numerical atmospheric models for the forecasting of severe weather associated with convective storms. Parameters have been derived from soundings in the proximity of large hail, tornadoes (including tornadoes over water: waterspouts) and thunderstorms in the Netherlands. 66,365 radiosonde soundings from six stations in and around the Netherlands between 1 Dec. 1975 to 31 Aug. 2003 were classified as being associated or not associated with these weather phenomena using observational data from voluntary observers, the Dutch National Meteorological Institute (KNMI) and lightning data from the U.K. Met. Office. It was found that instability as measured by the Lifted Index or CAPE and 0–6 km wind shear independently have considerable skill in distinguishing environments of large hail and of non-hail-producing thunderstorms. It was also found that CAPE released below 3 km above ground level is on average high near waterspouts and weak tornadoes that mostly occur with low shear in the lowest 1 km above the Earth's surface. On the other hand, low-level shear is strong in environments of stronger (F1 and F2) tornadoes and increases with increasing F-scale. This is consistent with the notion that stretching of pre-existing vertical vorticity is the most important mechanism for the formation of weak tornadoes while the tilting of vorticity is more important with stronger tornadoes. The presented results may assist forecasters to assess the likelihood of severe hail or tornadoes.

Keywords: Convective storm; Radiosonde; Tornado; Hail; The Netherlands

1. Introduction

Weather forecasters use various techniques to predict the occurrence of convective storms that produce thunder and lightning. Hereby, parameters deduced from radiosonde data and numerical model data often play an important role. Examples of such parameters are the lifted index (Galway, 1956), the *K*-index (George, 1960) and the Boyden index (Boyden, 1963). These parameters can be calculated either from observational data or forecast data from numerical atmospheric models. The skill of various forecast parameters as predictors of thunderstorms in the Netherlands has recently been studied by Haklander and van Delden (2003). As a follow-up of this study, the following problem is addressed:

How can radiosonde-derived data be used to forecast some of the potentially hazardous phenomena that accompany some convective storms: large hail and tornadoes?

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In contrast with the study by Haklander and van Delden (2003), we have not tested the forecast skill of a large set of existing parameters. Instead, a number of parameters have been selected that represent a single aspect of the atmospheric environment. So, for example, instead of testing the quality of the Significant Tornado Parameter (Thompson et al., 2002), we have considered the various building blocks of this parameter, which in this case includes parameters representing vertical wind shear, instability and the lifted condensation level. We think that this approach will more clearly reveal which processes are responsible for the severe weather phenomena. A number of studies using data from the United States have addressed approximately the same research question as that considered herein. These include the studies of Rasmussen and Blanchard (1998), Rasmussen (2003), Thompson et al. (2002), Craven et al. (2002), and Brooks and Craven (2002). In selecting the parameters to study, we have obviously been strongly influenced by those studies.

2. Theory

The subject of this study is the occurrence of severe weather in association with deep, moist convection (Doswell, 2001) that may be accompanied by thunder. Deep moist convection can be regarded as an instability: a flow perturbation that (initially) grows by means of positive feedback on itself while convective available potential energy (CAPE) is converted into kinetic energy. The concept of CAPE is based on the concept of a parcel of air that originates from some low atmospheric level and is lifted upward while it expands adiabatically. If it becomes positively buoyant with respect to its environment it will automatically accelerate upward.

A parcel has CAPE whenever it can become positively buoyant after being lifted by some process (if we neglect the contribution of perturbation pressure forces on buoyancy). The level at which the parcel becomes positively buoyant is called its level of free convection (LFC). It ultimately becomes colder than its environment at its equilibrium level (EL). Integrating the buoyancy force from the LFC to the EL gives us the work done by this force or, equivalently, the amount of convective available potential energy that is converted to kinetic energy of the parcel:

$$\mathrm{CAPE} = \int_{\mathrm{LFC}}^{\mathrm{EL}} B_{\mathrm{T}} \mathrm{d}z = \mathrm{g} \int_{\mathrm{LFC}}^{\mathrm{EL}} \frac{T_{\mathrm{v}}'}{\overline{T}_{\mathrm{v}}} \mathrm{d}z. \tag{1}$$

(e.g., Doswell, 2001).

Here, B_{T} is the thermal buoyancy, T_v' is the parcel's virtual temperature perturbation defined with respect to its environment and $\overline{T}_{\mathrm{v}}$ is the mean virtual temperature. g is the acceleration of gravity.

From the above, it follows that two requirements need to be met in order for deep, moist convection to occur:

1. the presence of CAPE
2. the presence of a forcing sufficiently large to release the CAPE.

In some cases, however, very little or no CAPE is present near convective storms. Such cases have, for example, been described in detail by Carbone (1982, 1983) and Forbes (1985). These cases were associated with strong vertical wind shear and tornadoes. Dynamic instabilities may have played more important roles in these convective storms than the release of CAPE.

2.1. Vertical wind shear, convective modes and severe weather

Wind shear has an important influence on deep convection as it causes dynamic pressure perturbations that can strongly influence the organization of storms.

Vertical wind shear is the derivative of the wind vector with height. It is often expressed as the magnitude of the vector difference between the horizontal winds at two specified altitudes or bulk shear. We will do so in this study as well. Among other parameters, the following have been considered in this study:

- 0–1 km bulk shear $= |\boldsymbol{v}_{1\ \mathrm{km}} - \boldsymbol{v}_{1010\ \mathrm{km}}|$
- 0–6 km bulk shear $= |\boldsymbol{v}_{6\ \mathrm{km}} - \boldsymbol{v}_{10\ \mathrm{km}}|$

Here, $\boldsymbol{v}_{10\ \mathrm{m}}$, $\boldsymbol{v}_{1\ \mathrm{km}}$ and $\boldsymbol{v}_{6\ \mathrm{km}}$ are the horizontal wind vectors, respectively, at 10 m, 1 km and 6 km above ground level (AGL).

When vertical wind shear is small, one can expect single or ordinary cells to form that have relatively short lifetimes (e.g., Byers and Braham, 1949) and do not frequently produce severe weather. When shear is larger, multicell storms may occur, which are more likely to produce severe weather. When vertical wind shear is large, supercells may form: storms that have a deep and persistent rotating updraft (Doswell and Burgess, 1993). In a study using proximity soundings, Doswell and Evans (2003) found that the median value of the surface to 6 km AGL bulk shear near supercells was slightly above 20 m/s. These storms generally have longer lifetimes than multicells and single cells, and are known for their capability to produce severe weather including large hail, tornadoes and severe wind gusts. An important reason for this is that supercells may contain very high vertical velocities within both updrafts and downdrafts. These may significantly exceed the values predicted by parcel theory (Weisman and Klemp, 1984) and are a result of dynamically induced vertical pressure gradients that develop as a result of the interaction of the updraft with the environmental vertical wind shear.

In addition to strong bulk shear, the presence of high storm-relative helicity (SRH) (e.g., Droegemeier et al., 1993) has been recognized as a factor favoring the formation of supercell storms that move to the right of the mean lower tropospheric wind. SRH is defined as

$$\mathrm{SRH} = -\int \boldsymbol{k} \cdot (\boldsymbol{v}_\mathrm{h} - \boldsymbol{c}) \times \frac{\partial \boldsymbol{v}_\mathrm{h}}{\partial z} \mathrm{d}z. \tag{2}$$

Here, $\boldsymbol{k}$, is the upward unit vector, $\boldsymbol{v}_\mathrm{h}$ is the horizontal wind vector and $\boldsymbol{c}$ the motion vector of the storm.

The integration in this formula may be over various height intervals. SRH integrated from the Earth's surface to 3 km AGL has been found to be high in supercell environments (Rasmussen and Blanchard, 1998). Its value strongly depends on the storm motion vector $\boldsymbol{c}$ that must be estimated if the actual motion vector of the storm is unknown. In this study, the ID-method developed by Bunkers et al. (2000) has been used for that purpose (see Appendix A).

2.2. Large hail

Hailstone growth starts when so-called hailstone embryos, small ice particles, enter a convective updraft. These collide with lighter supercooled liquid water droplets that are carried upward. Initially, the fall velocity of the hailstone with respect to the air is approximately compensated by the upward speed in the updraft. Upon collision, the droplets freeze and add to the size of the hailstone. Knight and Knight (2001) present two equations that can be combined to yield:

$$\frac{\mathrm{d}D}{\mathrm{d}t} = \mathit{eff} \cdot V_T \cdot r_l \cdot \frac{1}{2}\frac{\rho_a}{\rho_i} \tag{3}$$

Here, D is the hailstone diameter, *eff* is an accretion efficiency coefficient, that usually lies between 0 and 1 (Knight and Knight, 2001), V_T is the vertical speed of the hailstone with respect to the air and r_l is the liquid water mixing ratio. ρ_a and ρ_w are the densities of air and water, respectively.

This relation shows that the speed V_T of the hailstone with respect to the air and the liquid water content r_l are important for a rapid hail growth. The time t over which Eq. (3) should be integrated is just as important. V_T and t are both closely related to the updraft speed at the altitude at which hail formation takes place. Summarizing, large hail is favored by

1. large updraft speeds,
2. high liquid water content above the freezing level,
3. long storm duration.

Large updraft speeds can be expected in storms that develop in environments of large CAPE or in supercell storms. Supercell storms can occur with both low and high CAPE, but almost exclusively with strong vertical wind shear. Storm duration is strongly influenced by wind shear. Updrafts of single cell storms may often not live long enough for hailstones to grow very large. Multicells and supercells that are associated with moderate to strong vertical wind shear live longer and may therefore be more often associated with large hail. Based on this, we may a priori expect that environments supportive of storms producing large hail are characterized by high CAPE and strong wind shear.

2.3. Tornadoes

Davies-Jones et al. (2001) distinguish two types of tornadoes. Type 1 or mesocyclonic tornadoes form within a mesocyclone, a parent circulation that is a few kilometers in diameter. Type 2 or non-mesocyclonic tornadoes are not associated with a mesocyclonic circulation. They are thought to form by the rolling-up of a vortex sheet along a wind-shift line into individual vortices. Type 2 tornadoes are generally weaker than type 1 tornadoes.

A strong association has been found between tornadoes and low LFC heights (Davies, 2004). Low LFC heights imply that upward acceleration may be expected to start at low altitudes. This implies that strong vortex stretching can be expected near the surface, so that there is a strong amplification of vertical vorticity that favors tornadogenesis. This effect may be more adequately measured by the amount of CAPE released below 3 km AGL (Davies, 2002), here abbreviated as CAPE3km. Considering the most-unstable parcel, this quantity is defined as

$$\begin{aligned} \mathrm{MUCAPE3km} &= \int_{\mathrm{LFC}}^{\min(\mathrm{EL},\ 3000\ \mathrm{m})} B_\mathrm{T} \mathrm{d}z \quad && \text{where LFC} < 3000\ \text{m} \\ &= 0 && \text{where LFC} > 3000\ \text{m or no LFC} \end{aligned} \tag{4}$$

It has also been found that strong tornadoes are generally associated with a low lifted condensation level (LCL) (Brooks and Craven, 2002; Craven et al 2002; Rasmussen and Blanchard, 1998). A recent theory that explains why LCL height could be of importance for tornado formation focuses on the role of the rear-flank downdraft found in supercell storms. It has been found that large LCL heights are usually associated with a (strongly) negatively buoyant rear-flank downdraft that inhibits tornadogenesis (Markowski et al., 2002). Tornadoes would therefore be more likely in environments with low LCL heights.

2.3.1. Mesocyclonic tornadoes (type 1)

Although the formation of mesocyclonic tornadoes is not completely understood, it has been observed with Doppler radar that they form under mesocyclones that are strong at low altitudes above the surface, although this is not a guarantee that a tornado will form (e.g., Trapp, 1999). Nevertheless, an important question is how such a strong low-level mesocyclone can form. Rotunno and Klemp (1985) have identified two sources for updraft rotation:

1. tilting of streamwise horizontal vorticity originating from the storm's environment
2. tilting of streamwise horizontal vorticity created by the storm itself by baroclinic processes

They found that rotation at mid-levels is primarily associated with the tilting of environmental vorticity, while low-level rotation is caused by tilting of the streamwise horizontal vorticity created by the storm itself.

Recent studies have indicated that tornadic environments are often characterized by strong wind shear in the 0–1-km layer, which implies the presence of large horizontal vorticity (Brooks and Craven, 2002; Craven et al., 2002; Monteverdi et al., 2003). Other evidence of this includes the study by Rasmussen (2003) showing that storm-relative helicity (SRH, Eq. (2)) integrated up to 1 km AGL in the environment of the storm discriminates rather well between tornadic and non-tornadic supercells. This indicates that the tilting of environmental horizontal vorticity is important for the formation of tornadoes even though Rotunno and Klemp (1985) have suggested that the generation of vorticity within the storm itself is responsible for the low-level rotation.

2.3.2. Non-mesocyclonic tornadoes (type 2)

Non-mesocyclonic or type 2 tornadoes form along convergence boundaries like fronts, outflow boundaries or wind-shift lines. Along these boundaries, a quasi-vertical vortex sheet may exist that may break up into individual vortices as a result of a horizontal shearing instability (Wakimoto and Wilson, 1989). The vortices can be stretched by convective updrafts located over the boundary and subsequently develop into tornadoes. Additionally, tilting of environmental horizontal vorticity may also play a role in generating vertical vorticity as it does in mesocyclonic tornadoes (Davies-Jones et al, 2001).

3. Methodology and data sets

The goal of this study is to identify the values of sounding-derived atmospheric parameters in the neighborhood of severe convective weather events and thereby identify which physical processes are important for their formation. For this, we need data from radiosondes released in the proximity, both in space and time, of severe convective weather events. Firstly, we will describe the data sets that were used in this study.

The following severe convective weather events have been considered in this study.

- tornadoes, including waterspouts (i.e., tornadoes over water);
- hail having a diameter of 2.0 cm or more in its longest direction

Most observations of severe weather and radiosonde observations were available after 1975, so that we have decided to focus on that period. Data of severe weather was obtained from the monthly magazine *Weerspiegel* of the weather amateur organization Vereniging voor Weerkunde en Klimatologie (VWK, 1975–2003). Data from this source were available to us since December 1975, the year in which Dutch weather enthusiasts established the VWK, called "Werkgroep Weerkunde" at that time. The magazine *Weerspiegel* has been the primary source of severe weather data for this study. The time period considered in this study is 1-12-1975 to 31-08-2003 or 27 years and 9 months.

A few comments need to be made about the way amateur reports from *Weerspiegel* were incorporated in the data set used for this study. Some questions may arise about the quality of these reports. Firstly, some observations of tornadoes by amateurs may have been influenced by the wish of seeing a tornado rather than the actual observation itself. Therefore, we have been very critical with any mentioning of a tornado in the texts. Often, more details are provided by the observer than only the fact that a tornado was observed. Based on this contextual information, we have made an assessment of its credibility and decided whether or not to include the tornado in our data set. We have made a distinction between tornadoes that occurred over land and over a water surface. We will call the latter waterspouts.

Secondly, The F-scale classification (Fujita, 1971) has been applied rather crudely. In some cases, the section in *Weerspiegel* provided F-scale assessments. We have followed these in most cases where available. In a few cases, the written damage description or photo material did not match the F-scale estimate. After a discussion with one of the current editors of the tornado section, a handful of cases were reclassified. Nevertheless, the final classification was likely not always correct. In assessing the F-scale classification, differences–noted by, Dotzek et al.(2000)–in structural strength between average houses in the United States and stronger brick houses common in the Netherlands have been taken into account.

Finally, hailstone sizes have been taken literally from the texts in *Weerspiegel* magazine. The sizes of the stones are usually reported in centimeters both by the observers and may have occasionally been estimated, although it is the standard procedure of VWK observers to measure the diameter exactly. Hence, it is likely that these data contain some errors. Reports did usually not mention the method that was used to determine the hail size.

3.1. Radiosonde data

For this study, we have used radiosonde data from six stations in and around the Netherlands. Table 1 lists the radiosonde data that were used and Fig. 1 shows their location.

The data sets contained data on temperature, mixing ratio, wind speed and wind direction at the standard pressure levels of 1000, 925, 850, 700, 500, 400, 300, 250, 200, 150, 125 and 100 hPa and at so-called significant levels between the standard levels, as well as at the Earth's surface. Significant levels are extra levels with temperature and humidity and/or wind data so that the measured vertical profile of these variables can be reconstructed reasonably accurately by linearly interpolating between them. It is possible that some errors have been introduced by having data only at these selected levels. The first step was to interpolate (linearly with respect to height) the available data of temperature, mixing ratio, and u and v wind components at pressure levels spaced 1 hPa between the surface pressure and 100 hPa. Height data were interpolated assuming the hydrostatic equilibrium. Then, various shear-related parameters could directly be computed. The parcel ascent curve of the parcel having the highest θ_{ep} in the lowest 500 hPa was then computed. The virtual temperature correction (Doswell and Rasmussen, 1994) was applied to both the ascent temperature curves and the environmental temperature to obtain a more accurate estimate of the parcel's thermal buoyancy. Data of some soundings resulted in unphysical values for one or more parameters. Those have been completely discarded from the analysis. The remaining numbers of soundings are given in the right column of Table 1.

3.2. Lightning data

In order to be able to classify soundings as thundery, data on the occurrence of lightning have been used that originated from the U.K. Met Office's Arrival-Time Difference System (Lee, 1986; Holt et al., 2001). This system makes use of the fact that lightning strikes produce radio-waves called spherics, that move outward from the source at the speed of light in all directions. The system consists of seven stations located in the UK, Gibraltar and Cyprus that precisely record the time at which spheric signals reach the station. By comparing

Table 1
The soundings used in this study

WMO-ID and name of station	Period	Synoptic hours (GMT)	Number	Number used*
06210 Valkenburg	01-07-2002–20-11-2002	00, 06, 12, 18	402	375
06260 De Bilt	01-12-1975–27-04-1985	00, 12	32,532	31,372
	28-04-1985–30-06-2002	00, 06, 12, 18		
	21-11-2002–31-08-2003	00, 12		
06447 Uccle	01-01-1990–31-08-2003	00, 12	9832	9675
10200 Emden	01-07-1997–31-08-2003	00, 12	4403	4377
10304 Meppen	02-01-1990–27-06-2003	occasionally at 12	1201	1185
10410 Essen	01-12-1975–31-08-2003	00, 12	19,446	19,381
Total			67,816	66,365

Fig. 1. Map showing the locations where the radiosondes used in this study were released.

the difference in time that spherics were recorded, their source locations can be fixed with an accuracy of less than 10 km over west-central Europe. For this study, lightning recordings from 1 January 1990 to 31 December 1999 were used.

3.3. The definition of 'proximity'

A difficult question in any study that is based on proximity soundings is to define what can be considered to be the "proximity" of a certain meteorological event. This problem has been addressed among others by Darkow (1969), Brooks et al. (1994) and Rasmussen and Blanchard (1998). If the criterion of proximity is very strict, the sample set will consist of soundings that represent the event's environment rather well, but in low numbers. If the criterion is less strict, a large sample set will result that contains soundings of which some may not represent the storm's environment very well. The trick is to find some optimum in between. Definitions of proximity employed by the various authors differ strongly. Darkow (1969), for example, requires the sounding to be within 80 km of the event and released within the time frame 45 min before to 60 min after the event. A somewhat subjective extra requirement that the sounding had sampled the same air-mass as that which entered the storm's inflow was additionally applied. In contrast, Rasmussen and Blanchard (1998) allow the sounding to be released within 400 km of the event and in a time frame of 3 h before to 6 h after the event. An additional requirement is that soundings should be located within a 150°-wide sector directed upstream of the event using the boundary-layer mean wind.

The number of severe weather reports used in the current study was rather small, so that the proximity criterion was not chosen to be very strict in order to retain a reasonable number of soundings associated with each particular type of severe weather. A maximum distance of 100 km from the sounding was thought to be a reasonable balance between the number of soundings and their representativity.

To ascertain that the air-mass that was sampled closely resembles the air-mass in which the event took place, without having to inspect every sounding individually, a proximity criterion was defined with respect to the moving air that is sampled by the radiosonde. A complication is that air usually does not flow in the same direction and at the same speed at all altitudes in the atmosphere. Nevertheless, it seems reasonable to assume that a proximity criterion defined with respect to a (virtual) moving parcel is better than one defined with respect to the fixed location where the radiosonde was released. Therefore, the following criterion is used:

A sounding is considered to be associated with an event when the event occurred within 100 km of a point advected with the 0–3-km density-weighted mean wind from the sounding location (at t_0) within a time period starting 4 h before the sounding was released until 4 h after that time.

Here, t_0 is 30 min before the official time of the sounding, because the balloons are usually released some time before the official time in order to be completed at this time. So, for a 12 GMT sounding, $t_0 = 11{:}30$ GMT.

This criterion is obviously not perfect. It does not ascertain that the same air-mass was sampled by the sounding as that in which the event took place. It is also likely that some of the soundings were released within convective updrafts or in nearby compensating downdrafts. Additionally, air at low altitudes may in some cases have originated from convective outflows. These effects have likely altered the calculated parameters significantly and negatively affected the representativity of the soundings relative to the unperturbed environment.

All soundings have been put in one or more of the categories listed below based on whether they were proximity soundings of severe weather events as defined above. Occasionally, more than one sounding was found to be representative of a single event. The resulting numbers of soundings in each category are listed in Table 2. All soundings outside the period for which lightning detection data was available (1990–1999) that were not associated with large hail or tornadoes have been excluded from the analysis and put in the n/a category.

Table 2
Categorization of the soundings

Description	Abbreviation	Number of events	Number of associated soundings
Not associated with thunder, hail and tornadoes	NONE	n/a	27,349
Associated with			
Thunder	THUNDER	n/a	2045
Hail 2.0–2.9 cm [a]	HAIL<3 cm	78	52
Hail>2.9 cm [a]	HAIL>=3 cm	65	48
Waterspouts	WATERSPOUT	56	35
An F0 tornado	F0	36	25
An F1 tornado	F1	53	39
An F2 tornado	F2	8	6
An F1 or an F2 tornado [b]	F1+	61	45
Not associated with large hail, tornadoes and no lightning detection available. Not used in analysis	n/a	n/a	34,162

[a] This is the diameter of the largest hailstone that fell during a hailstorm. Where it was reported that the hailstones were not round, the diameter considered was that in the stone's longest direction (where this information was available).

[b] Because of the very low number of the F2 tornado soundings (6), the groups of F1 and F2 tornado soundings have been treated as one group (labeled F1+), except where parameter values differed significantly (at the 5% confidence level) between the two categories.

We have studied differences of the distributions of parameters between the various categories. Sometimes it was not possible to calculate the value of a particular parameter for a sounding. For example, when wind data above 3 km AGL were missing, the 0–6-km shear could not be calculated. If that same sounding had complete temperature and moisture data, the calculation of CAPE, for example, would still be possible. As a result, the calculations of statistics for each sounding category (presented in the next section) are generally based on different numbers of soundings for each parameter.

4. Results and discussion

4.1. Instability and hail

As was noted above, large hail requires strong updrafts. According to parcel theory, the vertical speed in updrafts is determined by the amount of CAPE that is released. Although that is a simplification of reality, the degree of instability and the upward speed in updrafts are probably quite closely related.

Fig. 2. shows the distributions of the CAPE calculated using the parcel with the highest equivalent

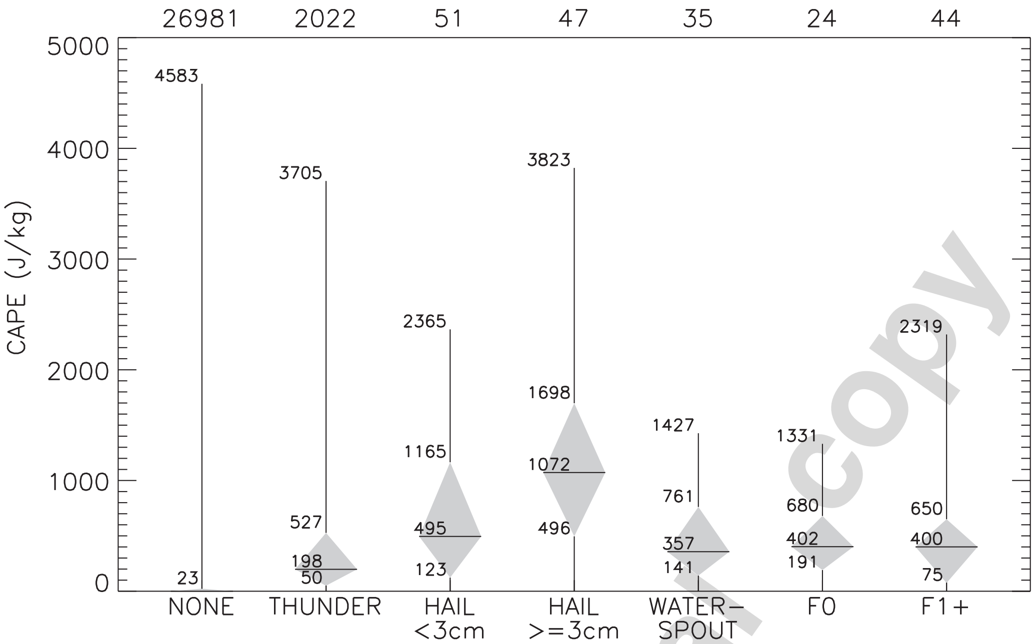

Fig. 2. Box-and-whiskers plots of the distribution of CAPE values of the various sounding categories. CAPE has been calculated for the parcel having the highest equivalent potential temperature in the surface–500 hPa layer. The figure shows the maximum (top cross) and the minimum (bottom cross) values. The box extends to the 25th and 75th percentiles and the whiskers to the maximum and minimum values. Numbers at the top denote the number of soundings in each category.

potential temperature (θ_{ep}) below the 500 hPa level with each sounding category. As expected, the CAPE is considerably larger with large hail events than with ordinary thunderstorms. The mean CAPE of both HAIL categories is significantly higher than the mean CAPE of the non-hail producing thunderstorms (at the 1% confidence level). This is likely a result of the fact that updraft speeds need to be stronger to sustain large hailstones than the speeds required for sufficient charge separation to cause lightning and thunder. The figures show that the CAPE is larger with the category of hail > 3.0 cm than with hail of 2.0–2.9 cm diameter, which is consistent with the idea that the stronger a convective updraft is, the larger are the hailstones that can be sustained.

4.2. Instability, LFC and tornadoes

From Fig. 2 it can be seen that the values of CAPE with waterspouts, weak (F0), and stronger (F1+) tornadoes is quite similar to the distribution with thunderstorms in general. CAPE is therefore not a very useful parameter to distinguish environments supportive of tornadoes from those that sustain thunderstorms.

MUCAPE3km (Fig. 3) seems to be a slightly more valuable parameter for tornado forecasting. This low-level CAPE parameter is significantly higher with waterspouts and weak (F0) tornadoes in comparison with thunderstorms. With stronger (F1+) tornadoes, however, the distribution is quite similar to that associated with thunderstorms on average. This finding may be explained by the theory that stretching of pre-existing vertical vorticity that is favored by high CAPE released at low altitudes is very important for the formation of weaker non-mesocyclonic tornadoes, while in contrast, the tilting of horizontal vorticity is more important for the formation of the stronger, mesocyclonic, tornadoes.

Fig. 4 shows the distribution of the height of the level of free convection of the most unstable parcel. The effect of tornadoes requiring the release of CAPE nearby the Earth's surface is probably reflected in this parameter as well: tornadoes appear to occur with low levels of free convection as these are associated with CAPE being released below 3 km AGL.

4.3. Wind shear and hail

From Fig. 5 it can be seen that a large fraction of the hail events <3.0 cm, and a majority of the hail events $\geq$3.0 cm is associated with 0–6 km shear below 15 m/s. It can be concluded that a part of the large hail events in the Netherlands, especially those $\geq$3.0 cm, is probably produced by multicell storms rather than supercells as the latter are usually associated with 0–6 km shear around or above 20 m/s (Doswell and Evans, 2003).

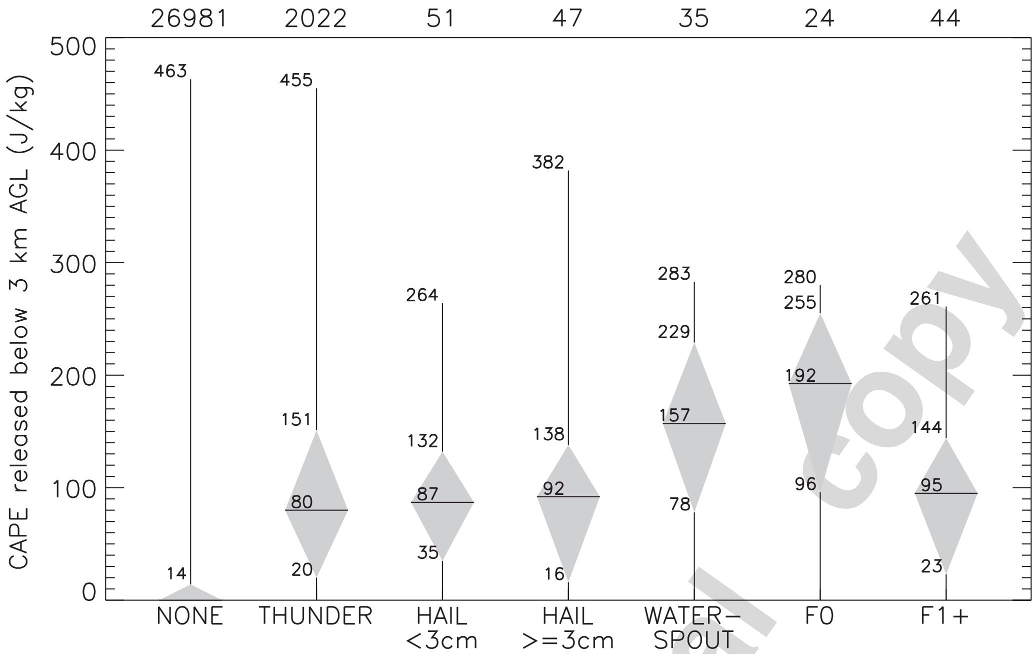

Fig. 3. As Fig. 2, but for the CAPE released below 3000 m AGL. CAPE has been calculated for the parcel having the highest equivalent potential temperature in the surface–500 hPa layer.

Hail <3.0 cm occurs on average with stronger deep-layer shear values than hail ≥3.0 cm. A possible explanation is given by the fact that hail ≥3.0 cm occurs almost exclusively from April to October, while smaller hail occurs year-round. Since thundery episodes in summer are on average associated with lower deep-layer shear values than those in winter, the hail events that are associated with the former will also be characterized by lower shear than the latter.

There are reasons to believe that hail in winter probably forms in convective updrafts that are often significantly enhanced by perturbation pressure effects, possibly including supercell or mini-supercell storms. Firstly, because the 0–6-km bulk shear

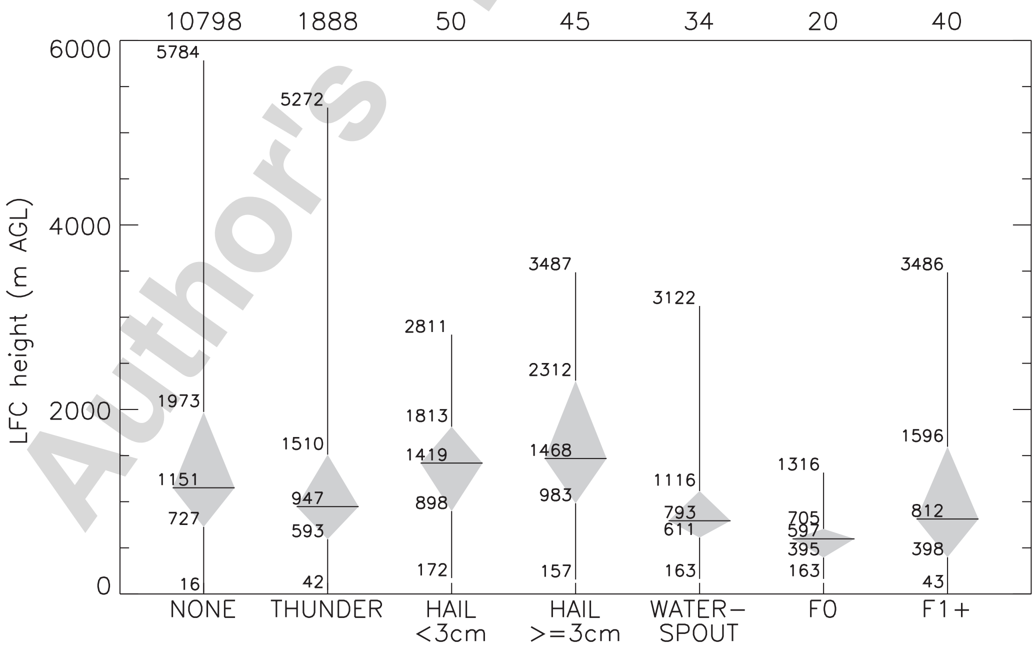

Fig. 4. As Fig. 2, but for the height of the level of free convection of the parcel having the highest equivalent potential temperature in the surface–500 hPa layer. Note that the number of soundings for which the LFC could be calculated is much lower than the total number of soundings as the LFC height is only defined for soundings that have nonzero CAPE.

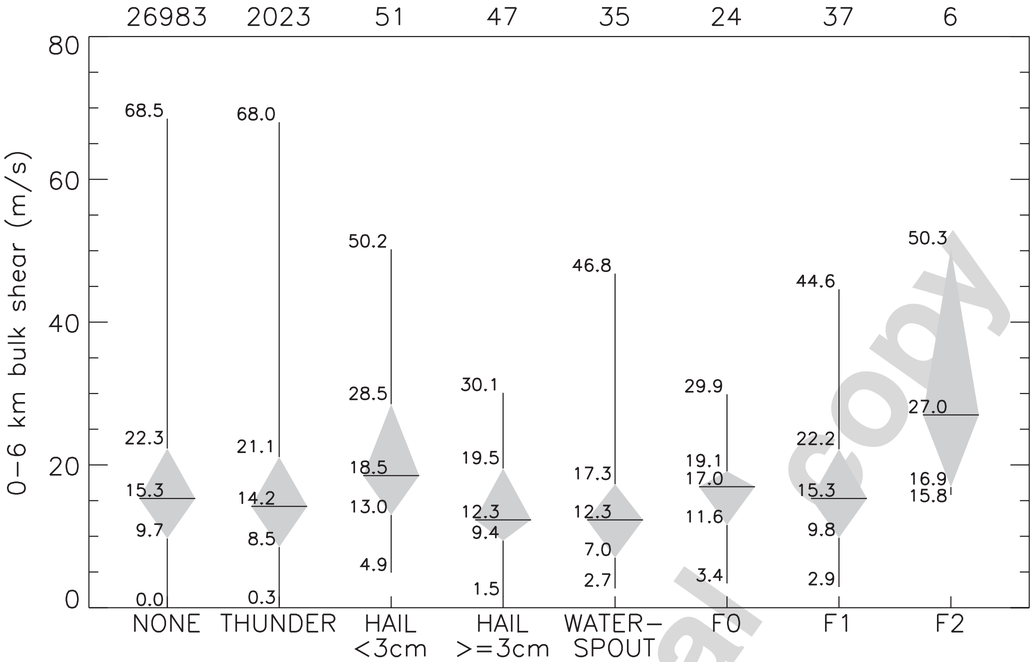

Fig. 5. As Fig. 2, but for 0–6 km bulk shear, i.e., the magnitude of the vector difference between the surface wind and the winds at 6 km AGL.

stronger in winter, pressure perturbations can be larger. Additionally, lower amounts of CAPE in winter suggest there is more need for a storm to require another forcing for upward motion than CAPE to sustain large hailstones (albeit somewhat smaller, i.e. 2.0–2.9 cm diameter).

In Fig. 6, the density of the large hail events in shear/instability space is normalized by the density of all events. It shows the percentage of the total number of events (i.e. thundery and non-thundery) per box that were associated with hail. The figure shows that the chance of finding a sounding to be associated with hail

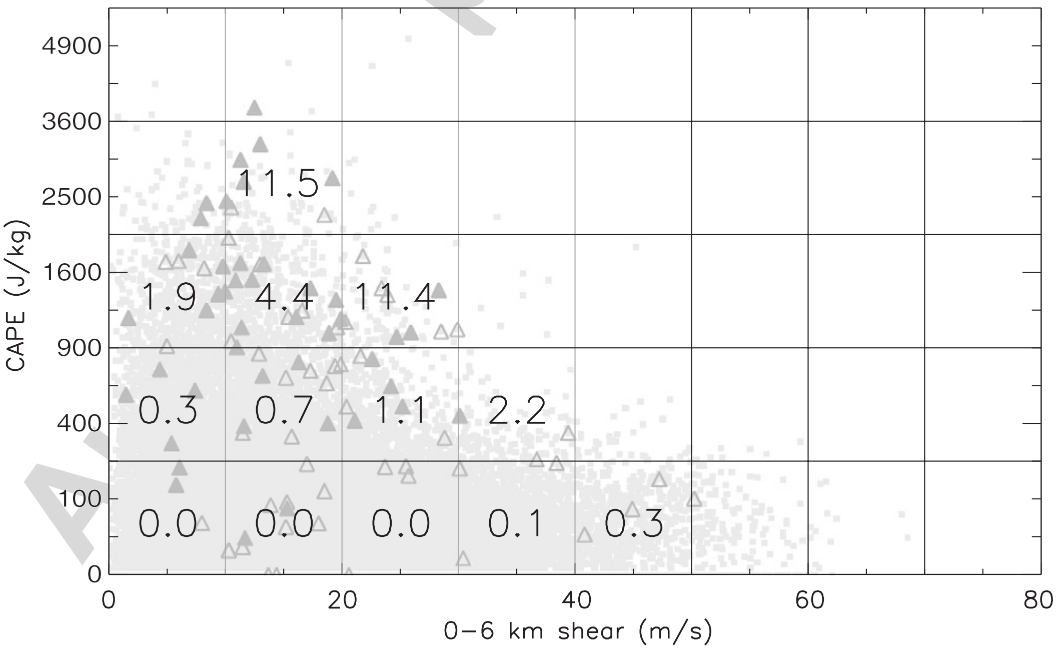

Fig. 6. Scatterplot of hail events with respect to 0–6 km bulk shear and CAPE calculated for the parcel having the highest equivalent potential temperature in the surface–500 hPa layer. Soundings associated with the HAIL 2.0–2.9 cm (51 soundings) and HAIL>3.0 cm (47 soundings) categories are represented by open and close triangles, respectively. Other soundings are represented by small grey squares (26,981 soundings). The numbers represent the percentages of soundings in the respective boxes that were associated with either of the two hail categories.

increases with increasing amounts of CAPE and shear. However, the typical large hail event is associated with shear in the 10–20-m/s range since combinations of high shear and high CAPE are relatively rare.

It is important to realize that Fig. 6 does not show the true ratio of hail events versus all events as the occurrence of large hail was (possibly strongly) underestimated: it is unlikely that all large hail that occurred was reported. Hence, the numbers in the figure should be interpreted in a qualitative sense. The true percentages could be much higher.

4.4. Bulk shear, storm-relative helicity and tornadoes

Fig. 7. demonstrates that the 0–1-km vertical wind shear with of F1 and F2 tornadoes is stronger than with F0 tornadoes and thunderstorms. With F0 tornadoes, the wind shear is weaker than the average values with thunderstorms. Possibly, the process responsible for their formation is inhibited by strong low-level wind shear and disrupts the formation of well-defined vortex sheets or the rolling-up into separate vortices.

In contrast to most F0 tornadoes, the F1 and F2 tornadoes that supposedly are in part associated with mesocyclones, seem to require strong 0–1 km wind shear to form. Low-level shear increases strongly with the intensity of the tornadoes. This suggests that the presence of strong low-level wind shear (implying that vorticity along a horizontal axis is present) is indeed important for the development of strong tornadoes.

The shear distribution of the waterspout events (of which no intensity estimates exist) reveals that this category consists primarily of the weaker, non-mesocyclonic tornadoes, although a few events are associated with shear values more typical of the stronger tornadoes.

The 0–6-km, or deep-layer shear (Fig. 5) is much less different between the F0 and F1 tornado categories than 0–1 km bulk shear. However, F2 tornadoes are associated with significantly (at the 1% confidence level) stronger shear than the F0 and F1 categories even though the sample size, 6 events, is very small.

In addition to the 0–1-km and 0–6-km bulk shear, storm-relative helicity was tested as a possible predictor for tornadoes. Box-and-whiskers plots of the storm-relative helicity in the 0–1-km AGL layers are presented in Fig. 8. The figure shows that both helicity parameters are high in environments of F1, and especially, with F2 tornadoes. For F0 tornadoes or hail, the distributions are centered just above 0 m^2/s^2 and are not significantly different from the NONE and THUNDER categories. As with 0–1 bulk shear, the highest values are observed with F2 tornadoes.

This may on first sight appear to solve the forecasting problem of strong tornadoes, but it should be recognized that the NONE category consists of many more events than the F2 category, that the absolute number of events occurring with, for example, more than 195 m^2/s^2 0–1 km SRH in the NONE category is much larger than that in the F2 category. This means that if one would decide to forecast tornadoes always

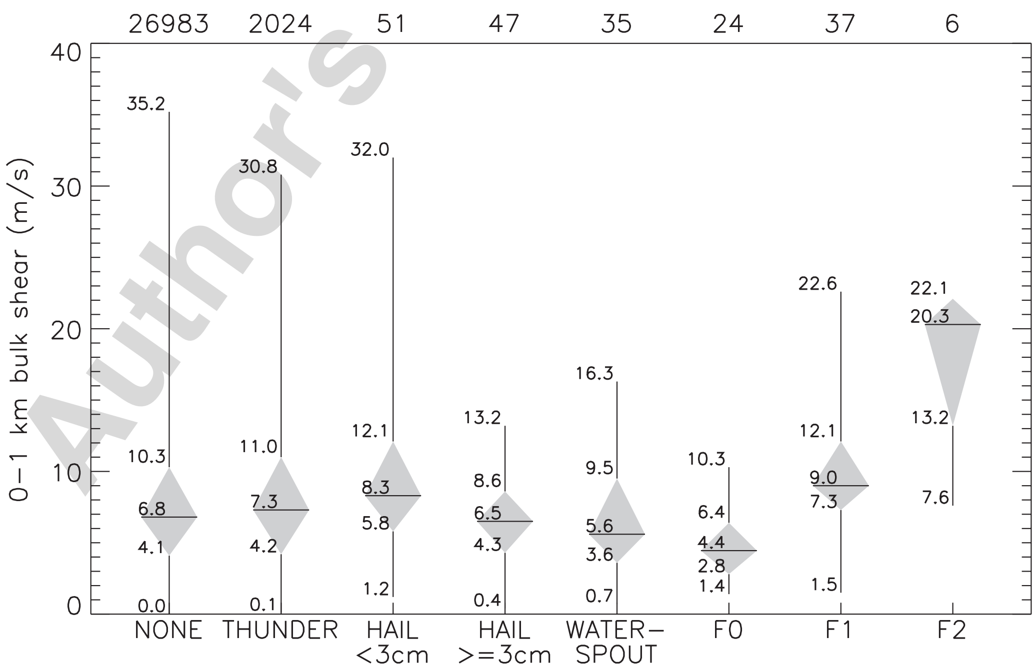

Fig. 7. As Fig. 2, but for the 0–1-km bulk shear, i.e., the magnitude of the vector difference between the surface wind and the winds at 1 km AGL.

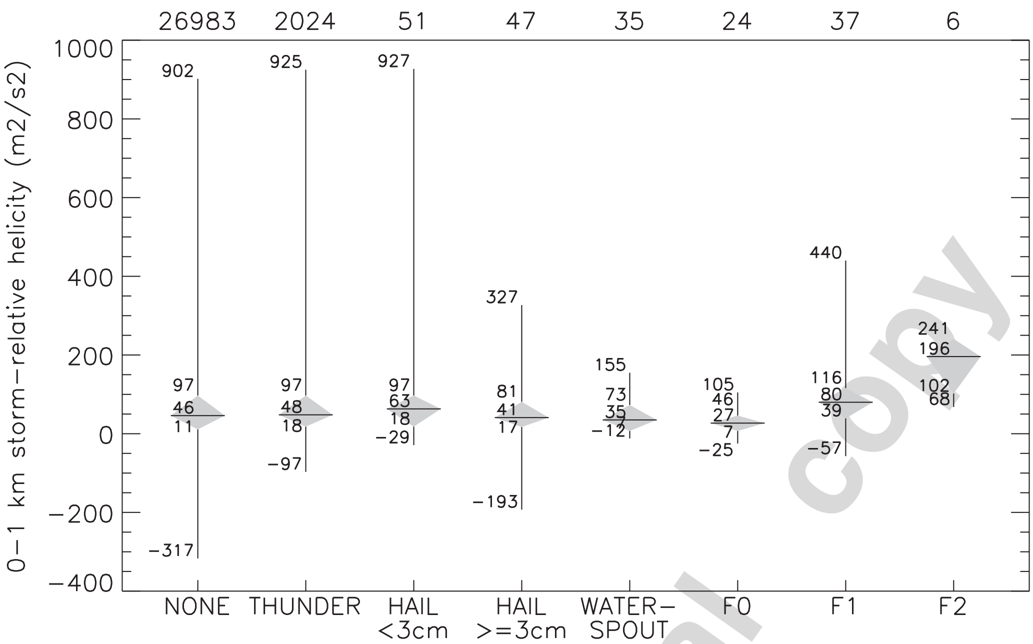

Fig. 8. As Fig. 2, but for 0–1 km storm-relative helicity.

when 0–1 km SRH in exceeds this threshold, this would result in a (likely unacceptable) high number of false alarms.

4.5. LCL height and tornadoes

The distributions of the height of the lifted condensation level (LCL) associated with the various categories is shown in Fig. 9. For the calculation of the LCL heights, a parcel is used consisting of a density-weighted mix of the lowest 500 m. It can be seen that the LCL heights are rather low with tornadoes: significantly lower than the situations of large hail, but there is no significant difference of LCL heights between the categories of waterspouts or weak tornadoes on one hand and thunderstorms on the other hand. Only the difference of the mean LCL height between F1+tornadoes and thunderstorms is statistically significant at the 1%

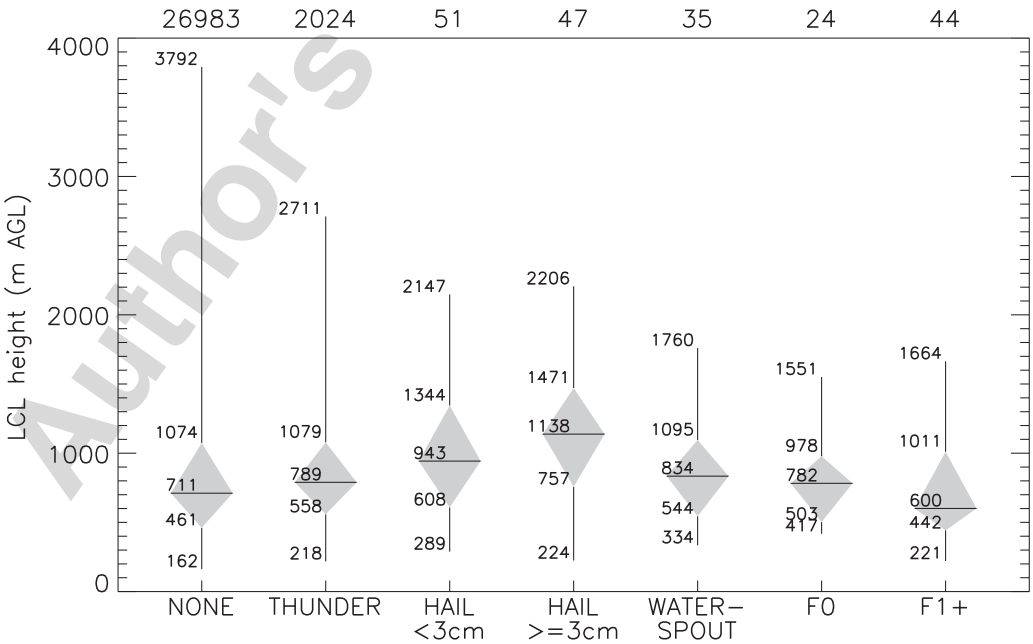

Fig. 9. As Fig. 2, but for the height of the lifted condensation level (LCL) of a parcel having the mean (density-weighted) moisture and temperature of the lowest 50 hPa of air above the Earth's surface.

confidence level. Nevertheless, the interquartile ranges between those categories show a very large overlap and the thunderstorm category is much larger. This means that in a given operational forecast situation this difference will be of little value.

The fact that LCL heights are not very useful in distinguishing tornadic environments from thunderstorm environments is in sharp contrast to what has been found in various U.S.-based studies (Craven et al., 2002; Brooks and Craven, 2002). This may be due to the fact that in the Netherlands thunderstorms often occur with low LCL heights, whereas in much of the United States, thunderstorms occur with LCL heights around or above 2000 m AGL. This means that LCL height is not as much a limiting factor for the development of tornadoes in the Netherlands as it is in much of the USA.

5. Conclusions

5.1. Main results

We have been able to establish a few significant relations between atmospheric parameters and the occurrence of hail and tornadoes.

Firstly, with respect to hail occurrence the following results were found. Large hail seems to be strongly associated with high values of CAPE—significantly higher at least than with thunderstorms on average. Especially summertime events that may have hail larger than 3.0 cm diameter occurs typically with moderately strong 0–6 km bulk shear, although the chance of hail does increase with increasing shear. However, situations of very large 0–6 km bulk shear are rare. Wintertime hail events are characterized by stronger (18.5 m/s) shear and lower instability, but also with smaller hail size (<3.0 cm).

Secondly, the following results were found in relation to tornadoes. Weak (F0) tornadoes occur with lower than average low level (0–1 km) bulk shear (average: 7.0 m/s, average F0: 4.4 m/s), while stronger tornadoes occur with much higher than average low-level bulk shear (average F1: 9.0 m/s , F2: 20.3 m/s). The 0–6 km bulk shear increases only, albeit strongly, when going from the F1 to the F2 categories (increasing from 15 to 27 m/s). This suggests that the weaker tornadoes are probably non-mesocyclonic, as the development of mesocyclones is associated with strong vertical wind shear. The stronger (F1 and F2) tornadoes occur with amounts of deep-layer shear supportive of mesocyclonic storms. Hence, some of them may be mesocyclonic tornadoes. A follow-up study could use radar data, where available, to confirm or reject the presence of a mesocyclone. Most waterspouts fit well into the category of F0 tornadoes and are probably of the same (non-mesocyclonic) type.

CAPE in tornadic cases is similar to that in any thunderstorm case, so that is not a useful parameter for forecasting tornadoes. The weak (F0) tornadoes are associated with high CAPE released below 3 km, typically 100–300 J/kg, but not the stronger tornadoes. All tornadic (weak or strong) environments are typically characterized by low LFC and LCL heights. Because the typical LFC and LCL heights with tornadoes do not differ from the typical values with thunderstorms in general, they have very limited value for forecasting tornadoes.

5.2. Limitations of this study, implications for operational forecasting

The relation between certain parameters and types of severe weather has only been established in a qualitative sense. This is due to the fact that only a small fraction of all severe events that actually occurred was reported, so that a quantitative estimate of the probability of severe weather as a function of the considered parameters cannot be made. Instead, only qualitative statements can be made, like "the chance of hail increases with increasing shear." The best way to use the above results is probably to monitor whether parameters are in the range supportive of a certain event type or are extreme in the context of climatology. This should contribute to the awareness that certain types of severe weather may occur, without directly warranting a warning to be issued to the public.

Another limitation of the presented results is that the analysis does not include other factors that influence the likelihood of severe weather than those that can be assessed using radiosonde soundings. Occasionally parameters may seem favorable for hail or tornadoes, but a convective storm–and hence, the resulting severe weather–does not develop at all. So the results presented should be used together with information that can tell a forecaster whether storms will form. Based on this and other factors a forecaster needs to consider (e.g., climatology, experience), a forecaster can decide whether a warning is indeed warranted.

5.3. Possible improvements to the methodology and suggestions for further research

In this study radiosonde data have been used as a proxy for the vertical wind, temperature and moisture profiles to be found in the direct environment of a severe weather event. Near-surface values of those variables are usually of high relevance to the degree of instability that can be realized, just as the surface wind is highly relevant for various shear parameters. Therefore, in any follow-up study, it is probably recommendable to derive more representative profiles by adjusting the original sounding data at low levels using surface observations nearby the events. This would likely lead to improved results.

Another way to obtain profiles more representative of those at the event location and time is to use numerical models of high resolution. This has been done already by Thompson et al. (2002, 2003) and may help to increase the size of the data set as numerical model data may be available at a higher temporal and spatial resolution than actual radiosonde soundings.

Future studies that use data of more severe weather observations may allow us to make quantitative estimates of severe weather probabilities instead of qualitative. For this to be possible it is crucial that more of the severe weather that occurs is reported and the data are stored in central location and is easily accessible. An initiative in this direction in Europe is currently being initiated under the name European Severe Weather Database (ESWD) (Groenemeijer et al., 2004). It makes use of the possibilities of the internet to share the information internationally and receive reports from weather amateurs that may be organized in national or regional networks.

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Appendix A. The ID-method (Bunkers et al., 2000)

We have here made the empirical assumption that the storm motion was equal to that which is given by the Internal Dynamics (ID) method (Bunkers et al., 2000):

$$c = \bar{v}_h + D \cdot \frac{(v_{6\ km} - v_{10\ m}) \times \hat{\boldsymbol{k}}}{|v_{6\ km} - v_{10\ m}|}$$

where $\bar{v}_h$ is the 0–6 km mean horizontal wind, $\boldsymbol{v}_{6\ \mathrm{km}}$ and $\boldsymbol{v}_{10\ \mathrm{m}}$ are the horizontal winds at 6 km and 10 m AGL, respectively, D is a constant of 7.5 m/s and $\hat{k}$ is the upward unit vector. This formula has been demonstrated to work well for supercells that move to the right of the main wind, which are the most common.

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