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Extreme Convective Windstorms: Current Understanding and Research

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Abstract

Convective windstorms are driven by downdrafts, the physics of which are relatively simple and correspondingly well understood. Damaging downdrafts are now often referred to as downbursts, and occur over a range of scales. Downburst also span a range of intensities, and even relatively mild convective windstorms can be a danger in some societal settings (e.g., recreational activities, aircraft operations, etc.). Moreover, convective windstorms can be isolated events or they can occur in large numbers within mesoscale convective systems. Those events which produced widespread damaging wind have recently come to be known as derechos.

Extreme convective wind events can be associated with either extreme intensity, or with large numbers of downbursts, or occasionally with both. Extremely intense events can attain windspeeds comparable to strong tornadoes: 75 m s^{-1} gusts are possible, with 50-60 m s^{-1} occurring virtually every convective season in the U.S. In derecho situations, damaging winds (exceeding 25-30 m s^{-1}) can occur over areas of up to 2000 km^2, and any one location might experience such winds for 30 min or more, with peak gusts approaching 60-70 m s^{-1}. The societal threat from such storms is comparable in terms of devastation to tropical cyclone events.

It recently has been suggested that some extreme wind events may be the result of supercells embedded within the mesoscale convective system. Often such storms are accompanied by large hail ($\geq$ 5 cm in diameter), which can increase the damaging effects of the strong winds. Recent research, including numerical cloud modeling and observations has suggested a consistent environmental structure associated with supercell-related windstorms. Thus, it appears that forecasting skill can be developed with such events.

1. Introduction

Convective wind events are a hazard to societies the world over, doing considerable damage and occasionally generating many casualties.[1] Most convection produces some straight-line wind as a result of outflow generated by the convective downdraft, and so anyone living in convection-prone areas of the world has experienced this phenomenon. On rare occasions, the intensity of the wind achieves the potential for doing damage. Whether or not damage actually occurs is the dependent on having structures in the path of the wind that can sustain damage. Although engineered structures typically are quite resistant to wind damage, many homes and outbuildings are quite vulnerable to damage from even relatively modest windstorms. In the United States, it is assumed that the potential for wind damage begins at around 25 m s^{-1} (50 kts). Of course, considerable damage occurs in situations where there was no anemometer, and so wind damage is graded according to its character: e.g., damage to tree limbs is considered non-severe, but uprooted trees is considered to represent a severe event. Figure 1 shows the recorded distribution of wind events considered to be severe in the United States.

Various human activities place people at risk from convective winds, notably aircraft operations and recreation. Most casualties from convective windstorms in the United States arise from such situations. Given the high vulnerability of aircraft operations during takeoff and landing procedures (the aircraft are operating on the margins of their flight

[1] In this discussion, winds associated with tornadoes and tropical cyclones are excluded from consideration.

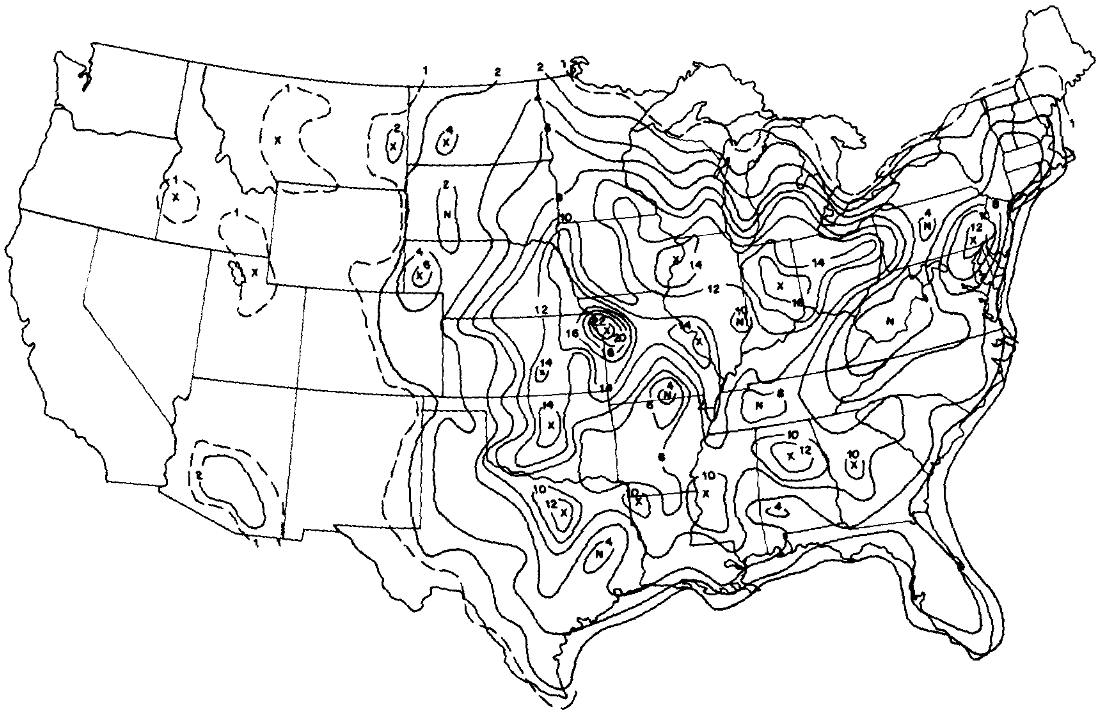

Figure 1. Contours of average annual frequency of reported severe convective wind gust events, per 10,000 mi^2 (25,900 km^2) for the period 1953-1980. Maxima are denoted by "X" and minima by "N" (from Doswell 1985).

"envelope" during such times), it does not take a particularly intense event from a meteorological standpoint to create many casualties. Commercial aircraft are less vulnerable than private aircraft, but their high occupancy means that rare events can have a large impact on casualty figures. Recreational boating also can account for many casualties in relatively modest windstorms, whereas most commercial craft are unlikely to be affected by marginal convective wind events.

In general, then, convective wind events and wind-vulnerable human situations are both common, whereas it is only infrequently that the two components of a convective wind-induced disaster are concatenated. In most such situations, the damage is confined in time and space because convective windstorms operate on time and space scales associated with single convective storms. The effected areas are only a few km^2 and the intense phase of the event is only a few minutes in duration. The opportunities for such space- and time-limited events to have a significant impact are few in number. On the other hand, occasionally, convective storms become prolific wind-producers, lasting for several hours and potentially affecting areas as large as 2000 km^2. These extreme events clearly have enormous damage and casualty potential. Although extreme events are, virtually by definition, rare events, their impact can approach that of tropical cyclones and certainly exceeds that of all but extreme tornadoes. Therefore, it appears that, with the growth of populations at risk from convective wind events, an effort to mitigate their casualty production is worth pursuing. Perhaps relatively little can be done to deal with the damage potential, except perhaps where building practices can be modified to make structures less vulnerable than they now are to low-end convective windstorms; it may not be cost-effective to build ordinary structures to withstand the rare extreme events.

2. Basic Physical Processes

Convective wind is the result of convective downdrafts, so to understand convective wind storms, one must understand the nature of downdrafts. This can be done with the aid of the vertical momentum equation,

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g, \tag{1}$$

where w is the vertical component of the wind (i.e., dz/dt, where z is the vertical coordinate), p is pressure, ρ is density, and g is the acceleration due to gravity. In this form, viscous forces have been ignored. Note that (d/dt) denotes the Lagrangian time derivative; i.e., following an air parcel. Suppose the variables are decomposed into a basic state, denoted by an overbar, and a perturbation, denoted by a prime, such that

$$w = \overline{w} + w', \quad \rho = \overline{\rho} + \rho', \quad p = \overline{p} + p', \tag{2}$$

and assume that the basic state is defined by the following,

$$\overline{w} \equiv 0, \quad \frac{\partial \overline{p}}{\partial z} = -\overline{\rho} g. \tag{3}$$

This decomposition also can be thought of as representing a parcel with perturbed properties embedded within a hydrostatically-balanced environment having no vertical motion. It is useful to observe that when the perturbations are small compared to the basic state,

$$\left(\frac{1}{\overline{\rho} + \rho'}\right) \approx \left(\frac{1}{\overline{\rho}}\right)\left(1 + \frac{\rho'}{\overline{\rho}}\right)^{-1}, \tag{4}$$

so that when (2)-(4) are used in (1), it can be shown that the parcel's vertical momentum equation becomes

$$\frac{dw'}{dt} = -\frac{1}{\overline{\rho}}\frac{\partial p'}{\partial z} - \frac{\rho'}{\overline{\rho}} g. \tag{5}$$

If the Ideal Gas Law ($p = \rho R T_v$) is used, where T_v is the *virtual* temperature (in °K), then it is easy to show from (5) that

$$B \equiv -\frac{\rho'}{\overline{\rho}} g = g\frac{\left(T_v - \overline{T}_v\right)}{\overline{T}_v}, \tag{6}$$

where B is used to denote buoyancy.[2] Note that the virtual correction usually is rather small and to a good approximation, it can be ignored when computing buoyancy. Finally, the effects of precipitation loading on the vertical motion are parameterized[3] by including a term in

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[2] In the demonstration, the contribution to the density from pressure perturbation is ignored. Note that buoyancy represents an unbalanced pressure gradient force.

[3] The presence of condensed water particles in the parcel is felt via viscous forces but these are on a scale that is not represented easily. Hence, the effect is treated as if it reduces buoyancy.

(5) that decreases buoyancy as the liquid water mixing ratio (ℓ) increases, leading to the final form of the parcel's momentum equation:

$$\frac{dw'}{dt} = -\frac{1}{\bar{\rho}}\frac{\partial p'}{\partial z} + B - g\ell \qquad (7)$$

This result contains the basic physical processes associated with the development of downdrafts (and updrafts, as well!). The first term in (7) is the effect of *perturbation pressure gradients* on vertical motion. In some storms, notably *supercells*, this term has a large effect on *updrafts* (Rotunno and Klemp 1982); however, there is not much reason to believe it has much of an impact on downdrafts, at least to a first approximation. Therefore, it will be ignored.

The second term is the effect of *buoyancy* on vertical motion. Clearly, in the case of downdrafts, we expect to find that B is negative: the parcel should be cooler than its environment. This cooling typically takes place as a result of phase changes (evaporation, melting, and sublimation). As shown by Kamburova and Ludlum (1966), precipitation particles that are small, but in large numbers, promote a maximum contribution to cooling and, hence, to creation of negative buoyancy. The major contribution to this process is from evaporation, so when I refer to "evaporation" in subsequent discussion, it should be understood that other phase changes can be involved, but that their contribution is usually minor.

The last term in (7) is that due to *water loading*. Whereas evaporation is promoted by large numbers of small droplets, it only takes a few large drops to contribute substantially to the downward acceleration of air parcels. This term is associated with storms having high precipitation rates. Comparing the effects of water loading to those associated with buoyancy, if a parcel has a liquid water mixing ratio of 1.0 g kg^{-1}, this is roughly equivalent to about 0.3˚K of negative buoyancy; alternatively, 1.0˚K of negative buoyancy is about the same as 3.0 g kg^{-1}; the latter is a large (but not extreme) value. Therefore, in general terms, negative buoyancy is typically the major contributor to downdrafts.

Finally, it should be observed that the contribution to negative vertical motion associated only with buoyancy (i.e., using pure "parcel theory") results in a prediction of the maximum downdraft of

$$-w_{max} = \sqrt{2 \cdot \mathrm{NAPE}}$$

where NAPE is the Negative Available Potential Energy,

$$\mathrm{NAPE} = -\int_{SFC}^{LFS} B\,dz\,, \qquad (8)$$

and where LFS denotes the Level of Free Sink for a descending parcel and SFC denotes the surface. This means that the maximum downward motion is associated with the integrated negative buoyancy. Even a relatively modest negative buoyancy can result in a substantial downdraft if it is maintained over a relatively large depth. A downward speed of 25 m s^{-1} results from the relatively modest NAPE value of 312.5 m^2 s^{-2}. To a first approximation, the maximum gust is roughly equal to the maximum downdraft speed.

3. Convective Storm-Scale Events

On the scale of a convective storm, downdrafts typically develop as storm-relative winds move precipitation out of the upper portions of the updraft, where it develops, and that precipitation then falls into relatively dry air. This process produces negative buoyancy through evaporative chilling. Also, the water loading effect contributes to downdrafts, as noted above. In the presence of storm-relative wind, which is present in most cases involving convection, the precipitation cascade region is somewhat removed from the updraft. Thi s precipitation cascade is roughly coincident with the downdraft (see Fig. 2).

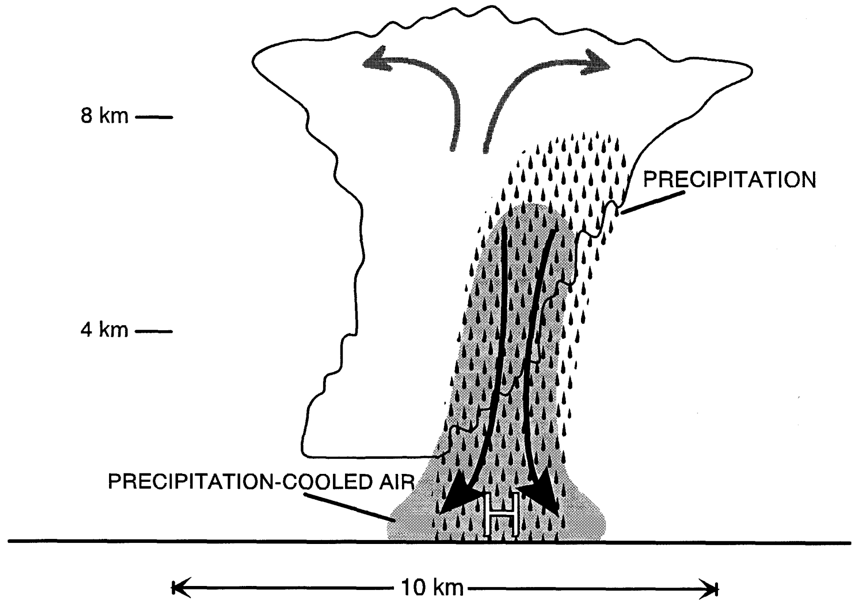

Figure 2. Schematic depiction of the mature stage of a convective cell, showing the precipitation (drop hatc hing) and the outflow (stippling).

Downdrafts lag the updrafts that produce them by about half a convective time scale (a useful value for which is the time needed for a parcel to rise from the condensation level to the e quilibrium level, or about 20 min). The duration of the maximum gusts with any individual downdraft is at most a few minutes, typically occurring shortly after the downdraft first r eaches the surface. Since the spatial scales of most convective cells are on the order of 10 k m, the size of the outflow during the time of maximum horizontal wind gusts is only a few k m. Such small-scale events have come to be called *microbursts* (Fujita 1978). It has becom e clear recently that microbursts, even when their intensity is below the arbitrary "severe" limit of 25 m s^{-1}, are still a considerable hazard to aviation (see Fujita and Byers 1977; Fujita and Caracena 1977, Caracena et al. 1989).

The so-called "dry" microburst event is one created by evaporation, but it is ultimately dr iven by the initial negative buoyancy. That is, evaporation produces the negative buoyancy, but the environments in which such events occur, namely with deep layers of nearly dr y adiabatic lapse rates and low relative humidities (Fig. 3) permit unimpeded descent of the chilled parcels. The circumstances leading to dry microbursts are relatively well-understood and they are fairly easy to forecast in the sense that the conditions in which they ar ise are fairly restricted (see, e.g., Wakimoto 1985; Caracena and Flueck 1988). On the othe r hand, the "wet" microburst appears to develop with some contribution from evaporation, and the negative buoyancy is maintained by continuing evaporation. Wet microbursts occur in situations with nearly moist adiabatic environments, having deep surface-based layers of high relative humidity. The key to developing strong downdrafts is the continuing evaporation of condensed water during descent to maintain saturation. Without such evaporation, the descending parcel would warm at the dry adiabatic rate and quickly los e its negative buoyancy. Water loading also may play a larger role than in dry microbursts. Much less is known about how wet microbursts arise and they are correspondingly more difficult to forecast than dry microbursts.

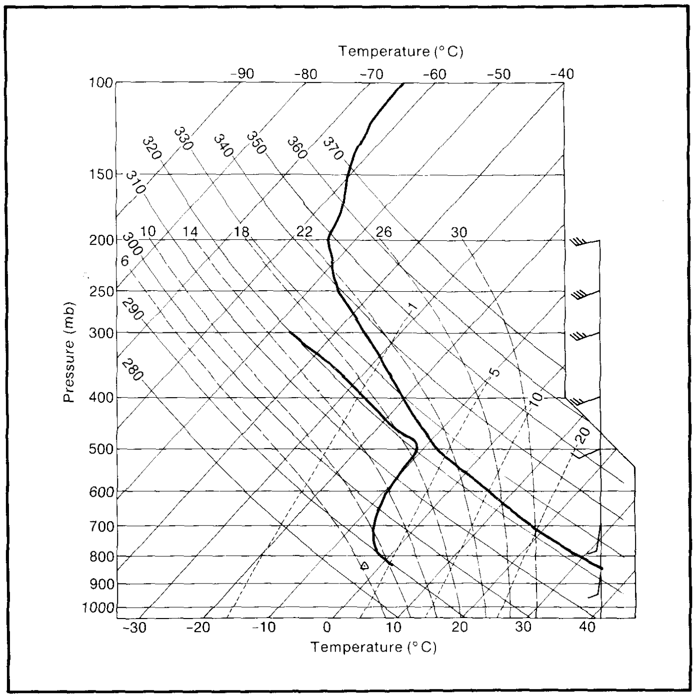

Figure 3. A composite of five 0000 UTC soundings at Denver, Colorado on days that produced dry microbursts on the Front Range area of Colorado (from Brown et al. 1982).

It is important to recognize that the conditions under which downdrafts are unstable do not necessarily coincide with those producing strong updrafts (Johns and Doswell 1993). This means that a convective event with comparatively weak updrafts still can produce stron g downdrafts. Some situations are capable of producing both strong updrafts and strong do wndrafts, which often occurs with supercells, but not all convective wind events are caused b y supercells. This means that severe and hazardous weather predictors keyed to locating co nditions for strong updrafts will not detect all cases involving downbursts.

Owing to the small space and time scales of isolated convective storms, it is unlikely that they can produce extreme wind events. Occasionally, a moderately intense wind event is as sociated with an isolated "pulse-like" convective storm, but meteorologically speaking, such events do not pose much of a threat of a disaster, except when they occur in combination with a particularly vulnerable human activity. Hence, we turn now to events associated with *systems* of convection.

4. Events Associated with Convective Systems

Conditions in which convection becomes organized into convective systems of various s izes and durations increase the chances for a convective wind-produced disaster, simply bec ause of the increasing time and space scale. The events producing the wind may not be sub stantially more intense than their more isolated counterparts, but the sheer number of them makes the chances for interacting with humans during their intense phases greater than with isolated storms.

In the United States, widespread convective wind events are sometimes referred to as *der echos* (Johns and Hirt 1987). Derechos invariably arise in association with mesoscale conv ective systems (MCSs). An MCS is composed of a number of individual convective cells, o ften arranged as a squall line of intense convection and a trailing "stratiform" precipitation area (Houze et al. 1989; Loehrer and Johnson 1993). As the convection evolves from the initial cells, new convective cells are initiated along the leading edge of the outflow from preceding cells and the system as a whole can maintain itself as long as sufficient moisture and lapse rates are present in the inflow. Note that convective systems of this sort can have a substantial inflow at low levels by virtue of their motion.

In these situations, it appears that a significant role is played by a system-relative, rear-to--front flow that begins in the lower mid-troposphere and descends to near the surface at the leading edge of the system, where the deep convective cells are. Although the origins of the rear inflow are not entirely clear, such a flow maintains a supply of subsaturated air to drive downdrafts by the evaporation of precipitation brought out of the convective region by a front-to-rear flow that begins ahead of the convective line and extends into the system's anvil region. These features are summarized in Fig. 4. Within such a system, the potential exists for numerous intense downdrafts, as the rear inflow interacts with the strong convection on the leading edge. As shown in Fig. 5, the damaging winds can be distributed over a large area.

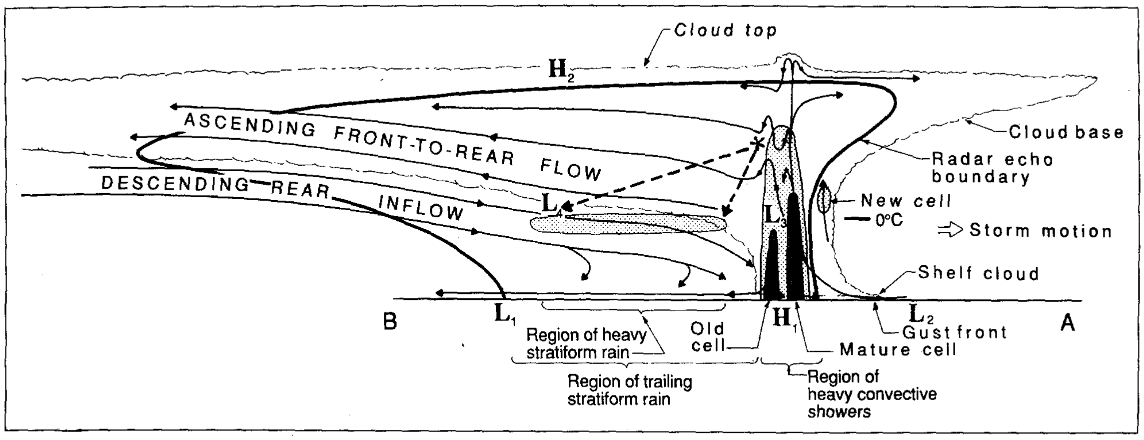

Figure 4. Conceptual model of a Mesoscale Convective System (MCS) with both a leading squall line and an area of trailing "stratiform" precipitation, shown in cross section parallel to the motion of the system (from Houze et al. 1989).

Many important windstorms are associated with bow-shaped radar echoes (Fujita 1978). These structures can arise on scales from 15 to over 150 km (Johns and Doswell 1993) and appear to arise from the development of strong rear-to-front wind flow. Weisman (1993) has suggested that the so-called "bookend" vortices (Fig. 6) often observed at opposite ends of a bow-shaped echo arise from tilting of ambient horizontal vorticity; his interpretation of their role is that they can enhance the rear-to-front flow between the vortices by as much as 30% through the development of favorable pressure gradients. This interpretation remains to be validated by observations, although the basic morphology of observed bow echo systems does, in fact, agree with the numerical simulations (Smull and Weisman 1993).

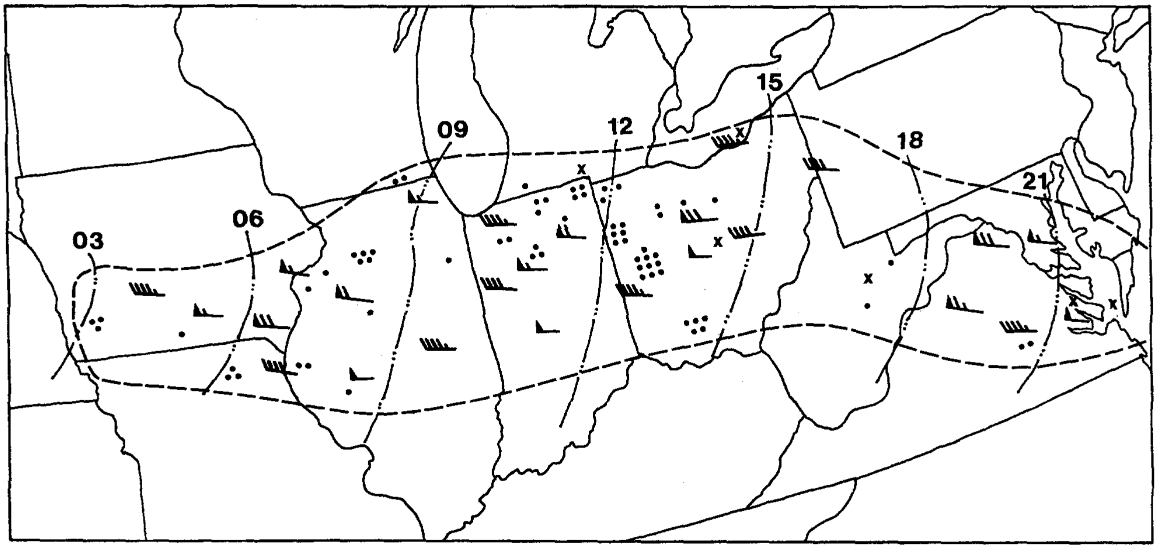

Figure 5. Damage swath produced by the derecho of 5 July 1980; three-hourly squall line positions are indicated by the dash-double dot lines, with UTC hours indicated. Officially measured wind gusts are indicated by the wind barbs, with a full barb indicating 5 m s^{-1} and a flag indicating 25 m s^{-1}. Personal injuries are indicated by dots and deaths with an "x."

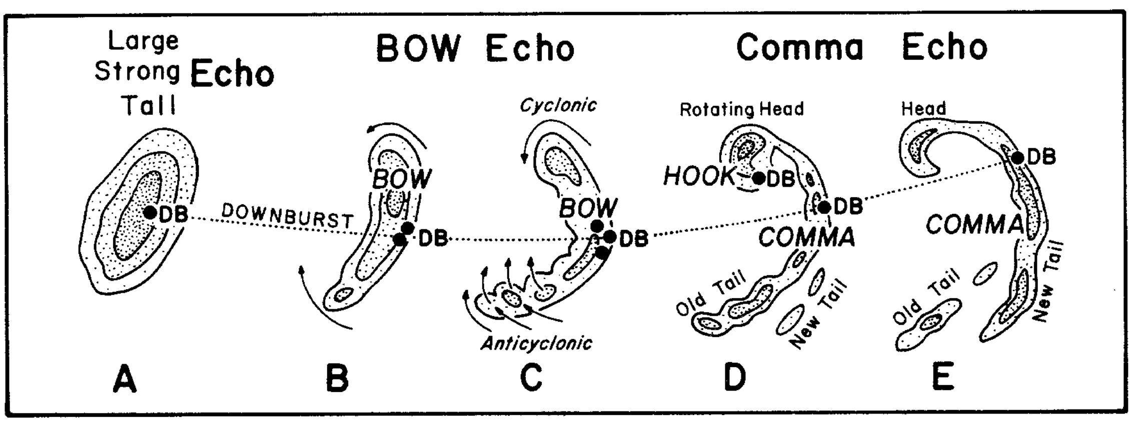

Figure 6. Schematic evolution of low-level radar reflectivity structure in a bow-shaped echo convective system (from Fujita 1978).

5. Supercell Events

It already has been noted that supercells sometimes arise in environments that exhibit the potential for both strong updrafts and downdrafts. This is a consequence of having high Convective Available Potential Energy (CAPE), where

$$\text{CAPE} \equiv \int_{LFC}^{EL} B dz\,, \tag{8}$$

with LFC denoting the Level of Free Convection and EL denoting the Equilibrium Level for an ascending parcel. This is the updraft equivalent of the NAPE. High values of CAPE are

invariably associated with (i) large values of low-level absolute humidity and (ii) low relative
humidity in the lower mid-troposphere. The latter condition allows the devel-
opment of step lapse rates in the lower mid-troposphere, which are necessary to develop hig
h CAPE (see Doswell et al. 1985). Clearly, a layer with low relative humidity and steep laps
e rates favors the development of downdrafts by the processes already described.

On occasions when supercells develop, they contain (by definition!) a mesocyclone, at le
ast in middle levels, if not near the surface. This paper is not the forum for a discussion of t
he origins of the supercell mesocyclone; interested readers should consult Davies-Jones an
d Brooks (1993) for a review. Recently, Brooks and Doswell (1993) have suggested that so
me supercells, which have considerable precipitation within their mesocyclones, thus becomi
ng what is known as High-Precipitation (or HP) supercells (see Moller et al. 1990), can dev
elop extremely intense and persistent surface winds. The existence of precipitation in the m
esocyclone may arise as a result of having a strong mesocyclone at mid-levels in the presenc
e of weak storm-relative flow at those levels (Brooks and Doswell 1993). In such a situatio
n, the precipitation would not be carried far from the updraft by storm-rela-
tive flow and the mesocyclonic circulation would entrain the nearby precipitation, thus wrap
ping large amounts of liquid water around to its rear, where it would encounter the dry, unst
able mid-tropospheric air. This evolution would favor the continuous production of evapora
tively-chilled air on the rear side of the mesocyclone and, as long as the mid-level mesocyclo
ne could survive being undercut by the low-level outflow, a quasi-continu-
ous strong outflow would develop. Given the downdraft instability and the persistence of s
uch events, a large swath of wind damage could result; these are potentially *extreme* con-
vective wind events, especially when the wind is accompanied by large hail, as has been obse
rved on occasion. Damage swaths of up to 2000 km^2 (or more!) are not inconceivable.

There are two lines of reasoning that support this hypothesis for how such storms arise.
First of all, such storms have been observed to arise (Smith 1993; Cummine et al 1992, Br
ooks and Doswell 1993) and the conditions in which they have been observed seem to fit thi
s model (Fig. 7). In the limited sample of events consider to this date, the CAPE values ass
ociated with such storms are quite comparable to those with classic tor-
nadic supercells, but the storm-relative wind speed profiles are quite distinct from tornadic st
orms and similar from case to case. The other form of supporting evidence is that a nu-
merical simulation using one of the environmental soundings from the cases produces the s
ort of "wall of wind" that is observed in these extreme events.

Given that the early results of this work continue to hold up to further scrutiny, the fact t
hat such storms arise in a reasonably limited form of environment, it should be possible to a
nticipate such storms. The extreme events do not happen often, but their large impact means
it would be valuable to be able to forecast the threat of such a potentially devastating convec
tive windstorm.

6. Summary and Discussion

Convective winds are a relatively common phenomenon, although most such events are
not very intense. On the scale of the convective storm itself, peak outflow winds have short t
ime and small space scales, reducing the threat to society, even when the peak winds become
relatively strong. However, convection often is organized into mesoscale systems in which
numerous convective cells develop, mature, and decay. In some situations, con-
vective systems can produce a large number of convective wind events, leading to a much lar
ger effected area than that associated with any individual cell. When storms influence a larg
e area, the chances for significant hazards increase. The majority of

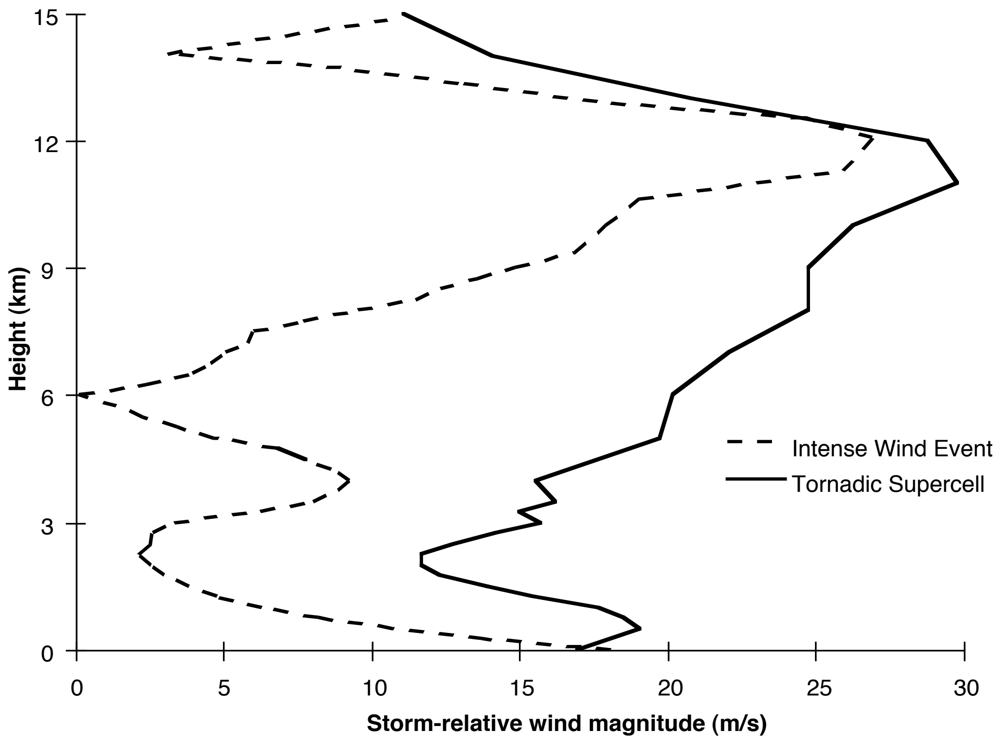

Figure 7. Storm-relative environmental wind speed profiles from an intense wind event (as described in the text) and a tornadic supercell, showing the weak storm-relative flow associated with the intense wind event.

windstorms in a convective system are of marginal severity, with only isolated events reaching high intensity. Nevertheless, the large area covered by such storms can result in major property losses.

The most threatening situation would be for a very intense convective wind event that also affected a large area. It appears that a few times each year in North America, extreme convective wind events of this sort do occur. To date, no such storm has struck a major city during a vulnerable time (e.g., the morning or evening rush hours). However, it is only a matter of time until this sort of unfortunate concatenation actually occurs. Given that the area affected can approach that of a tropical cyclone's damage swath, and certainly far exceeds that affected during a tornado outbreak (while not being as intense, of course), it is uncomfortable to imagine the potential devastation. When such storms are accompanied by large hail (e.g., $\geq$ 5 cm in diameter), the damage potential soars to even greater heights than when the wind occurs alone. The occurrence of hail has resulted in some of the costliest storms in United States history; coupling a fall of large hail with winds approaching 50 m s^{-1} could produce incredible damage in a populated area. Of course, economic losses to agriculture from such storms are already high, but do not attract much public attention, and such losses would be very difficult to mitigate with a 20-30 min warning.

A timely forecast may not be able to do much to mitigate the property loss, but could reduce the casualties. It appears possible to forecast these extreme events with some skill, but further research needs to be done to test the existing hypothesis about the interaction between the convective storm and its environment that produces the extensive swath of high winds.

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