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The role of vapor-water phase transitions in the generation of tornadoes

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Abstract. The possibility is considered of gaining a more accurate understanding of the generation of tornadoes through careful calculation of latent heat release during the vapor-water phase transition in a rotating stormcloud. The energetic characteristics for a tornadic cyclone and a tornado are compared. The relation between the size and power of a tornado is determined.

One of the most complex problems of modern meteorology is predicting particularly dangerous weather events such as tornadoes (known also as twisters). From the standpoint of geophysical hydrodynamics, the complexity of the problem is related to the essential three-dimensionality of air and the nontrivial structure of the velocity field in these atmospheric eddies.

A tornado is known to be the most intensive atmospheric eddy. According to some estimates, the wind speed in one may exceed 200 m/s [*Nalivkin*, 1984]. In the center of the tornado, the pressure drops sharply (10-15% lower than on its periphery): this forces the air to push toward the lower part. After reaching the edge of the center, the air sharply changes direction, and begins to spiral upward. As characteristic values of radius (R), tangential velocity (v) and vorticity ($\omega = \mathrm{rot}\mathbf{V} \approx 2v\ /R$) for a spiral tornado, the following values may be used: $R \approx 200$m, $v \approx 100$m/s, $\omega \approx 1\ s^{-1}$) (Figure 1).

Tornadoes are observed relatively rarely. For example, there are 5-10 cases annually in Russia. There is an approximately similar number in Western Europe. The exception is the USA, where recent data indicate that there are 700-800 tornadoes annually. The geography of tornadoes is extensive and nonuniform. Along with regions where tornadoes are encountered very frequently, there are areas where tornadoes never occur. The latter includes the polar regions and deserts. Clearly, such a distribution is related to the mechanism for tornadogenesis and, consequently, with the synoptic conditions in which this mechanism can be realized.

The typical synoptic conditions leading to tornadogenesis are associated with a polar front: the boundary between continental polar air, and warmer and moister marine tropical air. In this situation, very large horizontal contrasts in temperature are observed, most often along with a jet stream along the front. The polar front generally is unstable, and small perturbations in it often lead to the development of a cyclone. The warm moist air in the leading part of the cyclone will flow under the cold and significantly drier air. The resulting vertical temperature distribution proves to be highly unstable. Often a boundary, in the form of a thin inversion layer, which suppresses instability, exists between the warm air and the overlying cold air. If the inversion suddenly is disrupted in certain places, then the surface layer, which already is extremely unstable, displays a kind of explosion, and at these sites intensive convection begins. The air in the lower layer flows to sites of convective instability, forming extensive stormclouds. In such long-lived stormclouds we see formation of tornadoes.

Extensive factual material on clouds associated with tornadogenesis has been obtained by investigators from the Laboratory for the Study of Severe Storms and from the University of Oklahoma (USA), using Doppler radar [*Brandes*, 1977, 1981; *Lemon and Doswell*, 1979; *Ray*, 1976; *Biggerstaff et al.*, 1988]. The structure and evolution of a cloud from which a tornado forms has been studied in extreme detail. Initially, such a cloud is 10-20 km across and 10-17 km high. During the process of tornadogenesis, part of the enormous energy of the stormcloud is concentrated in an air volume with diameter of no more than several hundreds of meters, and height of not more than 6 km.

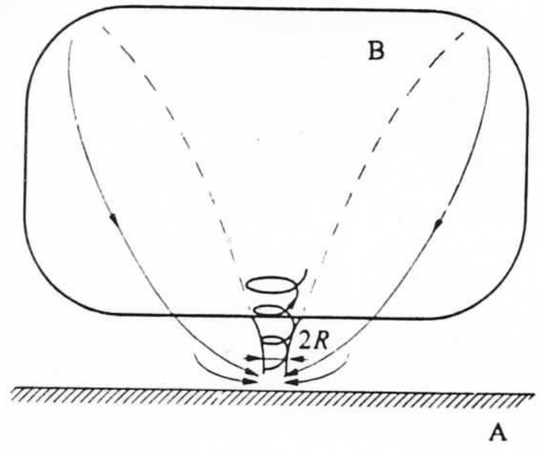

Figure 1. Diagram of a tornado. Letter A denotes the ground surface, letter B denotes a rotating stormcloud, within which there formed a tornado, depicted within by the broken line, and from the cloud base downward by the thick line; the thin lines with arrows show the circulation of air outside of and within the tornado; R is the radius of the tornado.

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Often, but not always, a forming tornado becomes visible (usually when the center of the tornado drops fairly close to the ground), appearing to the observer as a funnel-shaped cloud suspended from the parent cloud, and sometimes, unfortunately, reaching the ground). Visualization results both from condensation of water vapor present in the air, when it falls into the region of reduced pressure associated with the center of the tornado, and from the tornado carrying upward dust, rubbish, and debris. The vertical size of the funnel may range from several tens of meters to several kilometers, and its diameter where it adjoins the parent cloud may range from several meters to several hundred meters. Usually, the funnel resembles a cone tapering downward, although other types are found: thick columns, characteristic of very strong tornadoes (they are presumed to consist of several eddies of smaller size, rotating around a single center [*Bengtsson and Lighthill*, 1985]), or long cords that in some places stretch horizontally. Throughout the brief life (no more than several hours) of a tornado, the size and shape of the funnel may change sharply, reflecting the fluctuation in wind or change in characteristics of the inflowing air.

The powerful convective cell associated with a stormcloud ensures the indraft of air required for tornado formation, as well as being a necessary condition to keep the center of low pressure from filling with air from above. Tornadogenesis also requires that the rising air begin to rotate. The initial stage of this process, i.e., formation of a tornadic cyclone, can easily be understood from the vorticity equation for a homogeneous fluid

$$\frac{d\zeta}{dt} = \mathbf{k}\left(\frac{\partial \mathbf{v}_H}{\partial z} \times \nabla_H w\right) + \frac{\partial w}{\partial z}\zeta. \tag{1}$$

Here ζ is the vertical component of the vortex, $\mathbf{k}$ is the unit vector normal to the ground surface, v_H is the horizontal velocity, w is the vertical velocity, $\nabla_H = (\partial/\partial x,\ \partial/\partial y)$, z is the vertical coordinate, and t is time. Initially the vortex lines $(\partial v_H/\partial z)$ are horizontal. If there is a western shear flow, then they are pointed northward. When horizontal wind interacts with the rising flow $(\nabla_H w)$, the axis of rotation may tilt, so that a vertical vortex component will appear. *Barnes* [1970] hypothesized that this mechanism makes the major contribution to formation of an eddy storm. Further, the qualitative picture of the phenomenon proposed in this paper is thus: the rotating storm moves at an angle to the major flow, with the velocity growing with height. The movement of the stormcloud is slightly slower than the major flow, and points to the right (south) of it, into the region occupied by warm moist air. According to the Magnus effect* (*Because the flow has a vertical shear, and the velocity of the rotating storm in the lower layers is greater than that in the middle of the flow, direct application of the Magnus effect is not legitimate. It would be more appropriate to use the Bernoulli theory for a Beltrami flow, which represents with high accuracy the motion of air in a tornadic cyclone), the storm will move in this fashion with the condition that its rotation is cyclonic. Note that the overwhelming majority of tornadoes have cyclonic rotation. Observations using Doppler radar have shown that air rotation begins in the middle troposphere, at altitudes of around 4-8 km. First, the entire column of rising air with a diameter of approximately 10-20 km swirls. This stage is called a tornadic cyclone. After being generated at altitude, the rotation begins to be transferred to lower layers. This process results from the law of conservation of angular momentum for a ring of fluid particles compressed to the axis, which leads to extension of the vortex tubes

$$\left(\frac{\partial w}{\partial z}\zeta > 0\right)$$

and stretching of the "column." The intensity of the tornadic cyclone increases, and its diameter decreases to 2-6 km. A tornado usually forms near the center of a rotating tornadic cyclone in the region between rising and falling flows in the back part of a storm cell [*Lemon and Doswell*, 1979].

While the structure and fate of many tornadic cyclones have been sufficiently studied and tracked using Doppler radar, the actual tornado is not always found on radar screens. The characteristic size of a tornado is 50-300 m. The resolving capability of a Doppler radar depends on the distance between the radar and the studied vortex; for example, with a distance between them of approximately 15 km (which is not achieved by a single one of the existing networks of radar stations), the resolution is 300 m, which is on the border of resolution of even the large tornadoes. For this reason, our knowledge about the structure of an actual tornado is based primarily on random or indirect data. Except for the rare filmed materials, these are either untrained reports from surviving centerwitnesses, or remains from collapses, or unfounded speculations based on the first two sources of information. For this reason, notions about the wind field and pressure distribution in the center of the tornado are highly approximate. And for now, we are far from answering one of the major questions in the theory of tornadogenesis: "Does a tornado form from a given specific rotating stormcloud, or not?"

Very valuable in this regard are theoretical and numerical models, which can predict where and what should be measured, in order that a small number of measurements, still very rare and difficult to obtain, can be use to create a picture of the circulation in a tornado.

One of the very first attempts at theoretical explanation of formation of tornadoes was made by *Ferrel* [1889], who wrote that a tornado forms when a rising flow, inducing a buoyancy effect, encounters a "gyroscopic" wind field.

Currently, the total number of papers devoted to both theoretical study of tornadoes and to their numerical modeling is very large. A rather clear representation of the approaches to studying mechanisms of tornadogenesis and a more or less representative bibliography can be found in the paper by *Bengtsson and Lighthill* [1985].

Knowledge of these and certain recently published papers shows that the theory of tornadogenesis is far from complete, and that there is still no clear understanding of either the major causes for tornado generation or the major mechanisms responsible for development of this process. Despite the fact that many papers suggest entirely likely mechanisms for intensification of vorticity and formation of spiral flows, which have been the subject of rather detailed quantitative study, there remains a series of important unsolved problems. Nearly all models lead to the conclusion that the sizes and intensity of a tornado is dictated largely by processes in the boundary layer, and in particular depend heavily on the vertical velocity shear and the corresponding friction coefficient. The development of a tornado is described as occurring simultaneously in the entire air column extending from the ground to altitudes of 3,000-4,000 m (in the model, the boundary of the transition from the unstable stratification to the stable [*Mal'bakhov*, 1972; *Howells et al.*, 1988], while the major physical mechanism describing the explosive nature of growth in velocity is thought to be the transfer of momentum of the quantity of motion from the tornadic cyclone's periphery to the center by a radial current induced by processes in the boundary layer and depending definitively on the friction coefficient. Although at certain values for parameters this scenario facilitates modeling of an intensive vortex of small dimensions, it runs into a series of difficulties in explaining the behavior of actual tornadoes.

Thus, according to the models' purely dynamic description of the mechanism for growth in vorticity, the results suggest that tornadoes of smaller dimensions are more intensive. In this context, the vertical and tangential velocities in the direction from the center of the tornado to the periphery drop quite smoothly, while the process of tornado weakening due to friction occurs in stages. This does not agree completely with well known data about tornadoes, specifically: a) a tornado appears at the base of a rotating cloud and grows downward; b) the tornado is very compact; the border of transition from large velocities within the tornado to the surrounding air is estimated for many cases of interrupted pattern to be no more than one meter; c) more intensive tornadoes are large; d) the process of tornado weakening occurs dramatically, generally accompanied by a strong downpour.

This last fact, as well as analysis of the conditions for generation of tornadoes, prompt the conclusion that the major cause and primary mechanism responsible for the development of a tornado is the release of latent heat in the vapor-water phase transition. The majority of the models use a highly simplified description of the phase transition. Often, it is done simply by introducing, in some region of the tornadic cyclone, a non-zero buoyancy force into the equation for vertical velocity [*Nalivkin*, 1984].

Aside from general concepts, there is a series of facts we can cite to support the "moist" hypothesis, and to show that its use facilitates interpretation of some features of tornado behavior.

First we shall estimate the major types of energy in a tornadic cyclone and in a tornado: the kinetic energy

$$K = \frac{1}{2}\int_V \rho v^2 dv,$$

free energy

$$A = c_p \int_V \rho \frac{T'^2}{2T_{cp}} dv$$

[*Kurganskiy*, 1981], and that part produced during the phase transition of latent heat, which may turn into kinetic energy. The latter can be estimated using a method analogous to that used by *Golitsyn* [1980] in estimating the efficiency of convection from the transformation of heat, introduced into a fluid layer, into kinetic energy. Disregarding the molecular transfer of heat, and assuming that the flow of heat from the underlying surface is small in comparison with the flow arising in the vapor-water phase transition, we write the equation for the balance of heat flow at the z level for convection of moist air

$$\int_0^z \rho(-\mathcal{L}\frac{ds_m}{dt})dz' = \rho c_p \langle wT' \rangle, \tag{2}$$

where ρ is the air density, c_p is the specific heat capacity of dry air at constant pressure, $\mathcal{L}$ is specific heat of condensation, s_m is specific humidity of saturated air, angle brackets denote averaging along the horizontal, and T' is the deviation of temperature from the mean. Now, integrating (2) for height from 0 to the upper boundary of clouds d, we get

$$\int_0^d dz \int_0^z \rho(-\mathcal{L}\frac{ds_m}{dt})dz' = GH. \tag{3}$$

Here

$$H = c_p T/g \approx 32 \text{ км}, \; G = (g/T_{cp})\int_0^d \rho\langle wT' \rangle dz \; -$$

is the rate of generation of kinetic energy in a layer with thickness d. Then integrating the left side of (3) by parts, we will get

$$\int_{z_k}^d (d-z)\,\rho(-\mathcal{L}\frac{ds_m}{dt})dz = GH \tag{4}$$

(z_K is the condensation level). Assuming $\rho(-\mathcal{L}\frac{ds_m}{dt}) \geq 0$ and using the mean-value theorem, we get the estimate

$$d' \int_{z_k}^d \rho(-\mathcal{L}\frac{ds_m}{dt})dz = d'\mathcal{L}\frac{dm}{dt} = HG, \tag{5}$$

where dm/dt is the rate of formation of condensed moisture in a layer with thickness d over a unit area, $d' \leq d - z_k$, $\mathcal{L}\,dm/dt$ describes the speed of release of latent heat in a layer with thickness d during the phase transition. We write (5) in the form

$$G \approx \frac{d'}{H}\mathcal{L}\frac{dm}{dt}. \tag{6}$$

Clearly, only d'/H part of realized heat can turn into kinetic energy. Therefore, we shall consider the value

$$S = (d'/H)\int_V \mathcal{L}\rho_n dv$$

(ρ_n is the density of water vapor) to be an upper estimate of this kinetic energy obtained due to the liberation of latent heat.

As our characteristic sizes and velocities of a tornadic cyclone and a tornado, we use the following: diameter of the tornadic cyclone (d_u) is 10 km, its height (h_u) is 10 km, the velocity (v_u) is 10 m/s; these values for the tornado are, respectively, d_c --400 m, h_c --5 km, and v_c --100 m/s. We assume the water vapor to be found in the lower 1.5-kilometer layer. In estimating the free energy, we assume $T' \gg 4°C$, [*Riley and Bosart*, 1987]** (**The technical report of the Lithuanian Republic Administration for Hydrometeorology and Environmental Control on the tornado in the Vilnius region on May 10, 1985, kindly provided to us by M.A. Masterskikh, states that the horizontal temperature gradient in the tornado region was of the order of ≈8°C/100 km), i.e., we take that component of free energy that is associated with the horizontal heterogeneity of the temperature field, and responds to baroclinic instability. In this case, we get the following estimates:

Table 1

	Tornadic cyclone	Tornado
S	5×10^{14} J	10^{12} J
A	5×10^{12} J	2×10^{11} J
K	3×10^{13} J	3×10^{12} J

* J is Joule

This suggests, first, that a tornado is not an isolated object. Both during the process of its genesis and with further growth, it must be associated with a tornadic cyclone, since there is neither sufficient potential energy nor "available kinetic energy" contained within the tornadic volume to yield the enormous velocities observed within it. Second, it is clear that the free energy associated with the baroclinic instability of the atmosphere is not the major source of energy for the tornado: its reserves in the tornadic cyclone are comparable to the kinetic energy in a tornado, and it is unlikely that it would completely shift to energy of motion of the tornado. The table suggests that the major energy supply for the tornado is latent heat of water vapor. This makes the process of tornadogenesis essentially irreversible. With weakening of the tornado, the spent vapor that changed to condensed fluid pours out of the center with a strong downpour. From this precipitation we can also estimate the tornado's energy from the formula $K \approx \mathcal{L}\,\Delta m(d'/H)$, where Δm is the total quantity of precipitation from the tornado.

Note that the strong rain that has begun in the center of the tornado, carrying air with it, may sharply reduce the speed at which latent heat is released, and lead to the sudden collapse of the tornado. We can hypothesize here a mechanism to explain the sharp boundary of the tornado. With the enormous vertical velocities existing within the tornado's center resulting from condensation, drops--even those of extremely large size--will be found within the tornado, kept from falling by the Stoke's force. Moreover, because of the instability of large drops, the very large drops may start to fall and then break down into smaller drops and remain within the tornado. The large amount of condensed fluid may lead to a significant increase in the viscosity coefficient within the center. Because of this, the center of the tornado may be considered a solid, and the column of the transition layer may be estimated as the thickness of the boundary layer $\delta \sim l/(\mathrm{Re})^{1/2}$ (l is the characteristics size of the center, Re is the Reynolds number), which yields approximately 0.5 cm in the case of molecular viscosity, and approximately 5 m in the case of turbulent viscosity.

Now consider, from the standpoint of "moist tornadogenesis," the link between sizes of tornadoes and their energy. At the developed stage of a tornado, when the process can be considered steady, the rate of energy dissipation (a major characteristic of motion at all scales [*Kolmogorov*, 1942]), equals the rate of energy generation due to release of latent heat. Thus, assuming that the horizontal linear size of a tornado, l, depends on the speed of generation in it of kinetic energy, Q, and the characteristic speeds within the tornado, or its energy, E--which is the same thing, from the concepts of dimensionality we get

$$l \sim \frac{E^{3/2}}{|Q|}, \tag{7}$$

where $E = V^2/2$, $Q = -\kappa\mathcal{L}ds_m/dt \approx -\kappa\mathcal{L}w\partial s_m/\partial z$, ($\kappa \leq d'/H$ is the efficiency coefficient of transformation of latent heat into kinetic energy). We shall consider E and Q to be local characteristics, dependent on coordinates z, and hypothesize that κ is a weakly changing function of z and T. Then, since the vertical velocity in the tornado is

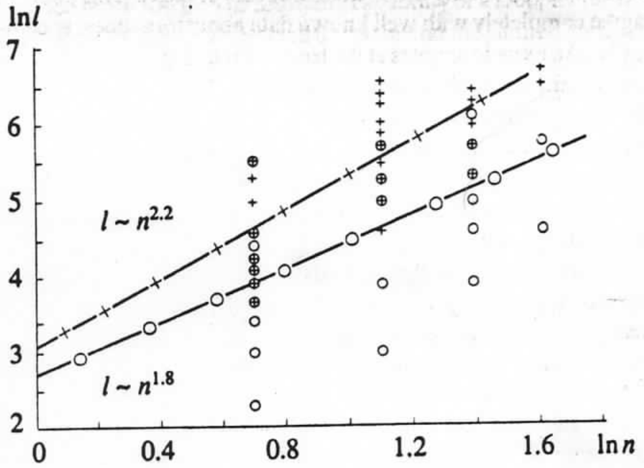

Figure 2. ln$l$ as a function of ln$n$. Crosses denote data relating intensity of tornado to its maximum size; circles --to minimum size. Upper line (-+-+-+) for which $l \sim n^{2.2}$, corresponds to maximum values of l; lower (-o-o-o), for which $l \sim n^{1.8}$, corresponds to minimum.

of the same order of magnitude as the horizontal, and the point value of the tornado n according to the Fujita scale [*Snitkovskiy*, 1987] is proportional to its speed, (7) can be written as

$$l(z) \sim \frac{1}{\kappa\mathcal{L}\left|\dfrac{\partial s_m}{\partial z}\right|}\,n^2. \tag{8}$$

Because $|\partial s_m/\partial z|$ is an exponential function that diminishes with height, (8) suggests that l depends on z. This dependence describes well the conical shape of the tornado. Moreover, the resulting shape of the cone is very similar to depictions of a tornado from drawings of eyewitnesses [*Nalivkin*, 1984].

Snitkovskiy [1987] presents, for a series of tornadoes, data on both the scale number and the sizes on the ground level. The minimum and maximum width of the trail were indicated as the size. We used these data to calculate the dependence between the sizes and intensity of tornadoes. Figure 2 presents two graphs depicting lnl as a function of lnn. The upper line was constructed for the maximum size of l, and the lower--for the minimum. The tangent of the angle of the line's slope to the abscissa corresponding to the index of the degree when n equals 2.2 for the upper graph, and 1.8 for the lower. Bearing in mind that, for lack of data, we did not include the coefficient at n^2 in formula (8), a coefficient that is dependent on temperature of the surrounding air and the moisture reserves in it, the agreement of actual data with the calculated formula can be considered very good.

As for the tornado structure, we can confidently assert that in its internal part, the motion of air approximates a spiral flow, where the vectors of speed and vorticity become nearly collinear. Because of this notion of spirality,

$$H = \int \mathbf{v}\vec{\omega}\,dV$$

is an integral hydrodynamic characteristic of tornadoes. In the internal part of these vortexes, the density of spirality $h = \mathbf{v}\,^*$ exceeds by 3-4 orders of magnitude the values observed under normal meteorological situations. Evaluating the density of spirality from the scheme presented above, we get

$$h \sim \frac{|Q|}{E^{1/2}}. \qquad (9)$$

Again assuming that the vertical and horizontal velocities are of the same order of magnitude, and expressing $\partial s_m/\partial z$ as the difference between the dry adiabatic γ_a and moist adiabatic γ'_a gradients of temperature, we write (9) as

$$h \sim \kappa c_p \left|\gamma_a - \gamma'_a\right| \sim 1\ \text{м}/\varsigma^2.$$

This shows that the density of spirality should not depend on the intensity and sizes of a tornado, but only on air temperature and moisture reserves in the atmosphere.

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