

The Physics of Tornadoes: Part 1

by Dwight E. Neuenschwander

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"We learned to do what only the student of nature learns, and that was to feel beauty. We never railed at the storms... To do so intensified human futility..."
—Chief Luther Standing Bear (1929)

If the biosphere ever reaches a state of thermal equilibrium, life will no longer exist. According to the second law of thermodynamics, we *must* have lack of equilibrium for irreversible processes—including life—to occur.

Out-of-equilibrium thermodynamics allows that conditions will occasionally swing to extremes. But life as we know it can survive only within a narrow range of parameters such as temperature and pressure. Here on our planet, which offers an oasis amid a universe of temperature extremes and much cosmic violence, what are the upper limits, constrained only by matters of physical principle, on the strength of an earthquake or the wind velocity of a cyclone?

The strongest winds on this planet occur within tornadoes. Not all whirlpools in the atmosphere are tornadoes. A funnel cloud that drops momentarily out of the clouds overhead, or a “dust devil” pirouetting across desert sands under clear skies, are not tornadoes. A working definition of a tornado typically requires a vortex (whirlpool) extending from a thunderstorm and touching the ground. In tornado production the wind speed, humidity, temperature, and pressure orchestrate an unusually violent event that is always fascinating and sometimes deadly (see Fig. 1).

About 750 tornadoes strike the United States annually. Wind speeds in a tornado vortex are difficult to measure directly. Early efforts used video footage of debris carried in the vortex. By knowing the distances involved and the time between frames, one computed the speed. Today Doppler radar makes real-time wind-speed measurements feasible, especially with mobile Doppler units that can achieve good resolution if one can park the truck sufficiently close to the action. (Leave the motor running!) Before Doppler measurements were widely available, a scale was worked out by meteorologist Theodore Fujita, who came to the University of Chicago from Japan in the 1950s. He can rightly be called the founder of modern tornado science. From him we have the vocabulary of “mesoscale systems” and “wall cloud.” He showed that thunderstorms can spawn multiple-vortex tornadoes. From the analysis of photos taken after the epic Palm Sunday Outbreak of April 3–4, 1974, when 148 tornadoes swept across 14 states and



Figure 1) An F5 tornado heading for the author's neighborhood in Piedmont, OK, 24 May 2011. Photo by the author's neighbor, Terry Harris, who ducked for cover immediately after taking this photo.

into Canada, Fujita ranked their strengths by the damage they produced and correlated each ranking to wind speeds in what is now called the Fujita scale. The Fujita rankings as normally applied go from F0 to F5. An F1 tornado, with wind speeds from 73 to 112 mph (the English units stick in the mind from media weather reports), will peel the shingles off roofs and overturn mobile homes. An F5 tornado, with wind speeds from 261 to 318 mph, will blast the bark off of tree trunks, rip a house off its foundation, and toss your refrigerator into the next county and your automobile into the next block. An F6 ranking exists (319–379 mph), but such atmospheric winds are so far considered inconceivable on Earth. The 2011 tornado season in North America saw several deadly tornadoes in Alabama, Missouri, and Oklahoma. Although tornadoes occur locally, they are spawned in the larger environments of continental weather and geography. In this article we examine the larger picture for context and discuss some mechanisms for tornado formation. No single model exists that describes accurately every tornadogenesis event.

For the past 25 years my family and I have lived in central Oklahoma, which lies within the peak of the tornado frequency distribution in “Tornado Alley,” the section of North America where tornadoes are seasonal visitations. It extends across the prairie states east of the Rocky Mountains and tapers into the upper Midwest and across the South (Fig. 2).

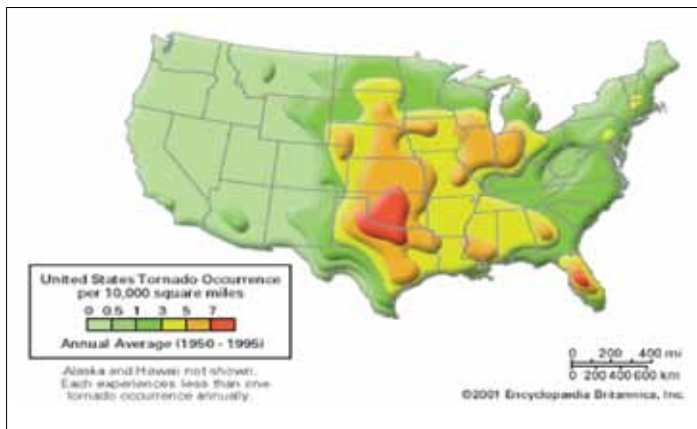


Fig. 2. Tornado geographic distribution (from *Encyclopedia Britannica*).

However, tornadoes can occur anywhere on Earth, and some of the more violent ones have happened outside the Alley. They form in a brew of physics and chance, and the damage and casualties they produce are catastrophic and sometimes tragic. At least they are impersonal acts of nature and not acts of vengeance; when we are hit by one we merely happen to be in its mindless way. Tornadoes are one of Nature’s ways of reminding us that the Earth does not need us but we need the Earth. About this point of respect and awareness, Chief Luther Standing Bear wisely observed in 1929,

We learned to do what only the student of nature learns, and that was to feel beauty. We never railed at the storms, the furious winds, and the biting frosts and snows. To do so intensified human futility, so whatever came we adjusted ourselves . . . without complaint.

Tornado damage is a price we pay for our civilization that values houses and infrastructure and settled communities. As urban sprawl expands to cover the Earth, the frequency of collisions between tornadoes and humanity’s developments will only increase (not to mention our society’s addiction to its carbon-fueled economy and our priorities of convenience that daily increase the supply of tornado fuel through global warming). Since we cannot do anything about the existence of tornadoes and because the conditions that make them possible also make life possible, we may as well understand them as best we can and prepare for their inevitable appearances. As Chief Luther Standing Bear said, the student of nature learns to feel the beauty. Tornadoes bring calamity but they are also fascinating.

Why does Tornado Alley exist? It is a matter of geography. To its south lies the Gulf of Mexico. To the north lies the Canadian prairie. We who live in the Great Plains joke how nothing stands between Texas and the North Pole but a barbed wire fence, and it’s down. But the joke illustrates the point: If

a chain of east-west mountains were to conveniently spring up as an offshoot of the Rockies, cutting across eastern Colorado, the Oklahoma-Kansas border, and into Missouri, it would break up the collisions that occur between the warm moist air moving north from the Gulf and the drier colder air moving south out of Saskatchewan.

In addition, to the west of Tornado Alley stand the Rocky Mountains. Air masses moving eastward over the Rockies are lifted to higher elevations by the terrain. Because air conducts heat poorly, as it raises into regions of diminished pressure it expands adiabatically. When the temperature drops sufficiently to precipitate water vapor, the air dumps its moisture onto the mountains. The dried air then sweeps over the lee side of the Rockies and moves into the prairie states, colliding with warm humid air coming up from the Gulf. From April through June, when about half of all violent tornadoes in the United States occur, these collisions carry optimal conditions for tornado formation. About one-third of Oklahoma’s tornadoes occur during one-twelfth of the year, in the month of May.

Tornadoes are small and states are big, so you can live in a tornado-prone region your entire life and never experience a direct hit personally. On the other hand, if you live here several years, you *will* experience numerous tornado warnings, acquire at least one near-miss story, and know someone personally whose house got clobbered. Let us do a rough estimate of probability. On May 24, 2011, an F5 tornado, about half a mile in diameter, cut a 65-mile swath southwest to northeast through central Oklahoma, thus hitting directly some 32 mi². The area of Oklahoma is about 70,000 mi² (Fig. 3).



Fig. 3. The path of the F5 tornado of May 24, 2011.

From Fig. 2 we may guess, for order-of-magnitude accuracy, that central Oklahoma, which lies in the peak of the tornado frequency distribution, occupies about half of the state. Thus if you were in this region on that day and the tornado was to set down at some random place within it, the probability of receiving a direct hit was about $32/35,000 \approx 10^{-3}$ —one in a thousand. About 2% of tornadoes are in the F4 or F5 categories, so let us say that about 14 such tornadoes occur annually in the United States. If one of them occurs in central Oklahoma and if you live here, statistically you will be hit by one every 1000 years. To say

it another way, for every 1000 families you know, about one per year will be hit.[1]

While the wreckage and ruin that tornadoes produce is ugly, an organized phenomenon in nature which quickly produces high-entropy disorder, they are awesome and beautiful in their physics. Their complexity arises from out-of-equilibrium thermodynamics in a multiphase fluid. The weather in general forms a nonlinear system—recall that “chaos theory” and the “butterfly effect” emerged from the study of meteorology.[2] In the Appendix (see www.spsobserver.org/tornadoes) I offer a glimpse of the tangled task faced by meteorologists who make mathematical models of weather. Supercomputers are necessary for them to step through the equations numerically. Even so, due to nonlinearities, weather forecasting can make useful predictions only a few days in advance, with uncertainties rapidly increasing the farther one tries to look ahead.

A Day with a Thunderstorm

Tornadoes are typically spawned in thunderstorms. The most prolific tornado nurseries are the county-size thunderstorms called “supercells.” To better understand supercells, let us start with a run-of-the-mill thunderstorm. Later we can see how its development gets altered into a supercell, leading to a possible mesocyclone and then a tornado.

The mechanisms of thunderstorm development were first documented in the Thunderstorm Project carried out by meteorologist Horace Byers and his graduate student Roscoe Braham of the University of Chicago. (It was Byers who brought Theodore Fujita to the University of Chicago.) They studied thunderstorms near Orlando, Florida, in the summer of 1946 and in Ohio the next year. Even though the project was motivated by thunderstorm-caused airplane crashes, these researchers somehow convinced pilots to fly instrument-laden aircraft through such storms. The results were published in Braham’s master’s thesis and in a now-famous 1949 volume by both authors.[3]

They documented how a thunderstorm develops through three phases. During the first “cumulus” stage a warm bubble of air is lifted upward. The lifting can begin when air flows up a mountainside or hitches a ride aboard a “thermal.” Solar heating of the landscape routinely produces convective updrafts that lead to cloud formation without thunderstorms, but thunderstorms need more vigorous lifting. A common lifting mechanism on the prairie that leads to severe weather occurs when a cold front or dry line moves across the landscape, making the warm, moist air slide up over the drier, cooler air.

Whatever the lifting mechanism, as the bubble rises to higher altitudes where the ambient pressure drops, the air inside the bubble expands and cools adiabatically. Moisture precipitates out and the latent heat released warms the bubble. Should this lifting and latent heating make the bubble buoyant, it accelerates upward while cooler air rushing in below drives further convection. Such buoyancy carries air upward at speeds that can reach 30 mph or more. Several kilometers aloft, if not before, the temperature inversion in the stratosphere above kills the updraft and the top of the cloud spreads out into an “anvil.”

As cloud droplets coalesce into raindrops, they begin to fall. The cloud now has a downdraft of less buoyant cool air and rain, in addition to the original updraft. This forms the second, mature “cumulonimbus” stage of the storm.

When the downdraft hits the ground and spreads out, the observer on the surface below feels the cool “gust front.” Should the gust front be sufficiently strong, it lifts some of the warm moist surface air it encounters, possibly propagating a new thunderstorm.

Back in the collapsing storm, if little horizontal wind exists aloft, the storm stands upright at altitude so that the rain descends through the updraft. Some of the drops are carried aloft again to freeze, perhaps looping down and up repeatedly, on each trip upward acquiring a coating of supercooled water which immediately solidifies, growing large hailstones. When a hailstone’s weight can no longer be supported by the updraft, gravity wins and the hailstone whacks your car at terminal velocity, sending you scrambling for covered parking—and later to a windshield shop.

If there is little vertical wind shear (horizontal components of velocity that increase in speed and change direction with increasing height), the rain-loaded downdraft eventually quenches the updraft. A thunderstorm may also be shut down if the gust front pushes away the incoming airflow that feeds the convection. In such ways does a thunderstorm enter the third “dissipative” stage as the updraft collapses and the cloud rains itself out.

A thunderstorm turns into a supercell when first, the rising air mass contains enough moisture whose latent heat drives an especially vigorous convection, which packs enough kinetic energy to possibly break into the stratosphere before forming the anvil; second, it encounters sufficient horizontal winds at high altitude to tip the cumulous cloud over so the downdraft falls outside the updraft; and third, enough vertical shear exists aloft to generate vorticity and build a “mesocyclone,” a large circulation in the storm that makes the storm chasers run for their trucks.

A mesocyclone might subsequently “tornado” through several scenarios that can present themselves in a supercell environment, although none are understood in detail. Nevertheless, the conditions that favor supercell formation can be seen converging in advance and alerts issued to the public. For instance, last May 21, a Saturday, the Oklahoma City television meteorologists started saying, with emphasis, “Next Tuesday appears to be primed for severe weather, with high probability for thunderstorms, damaging winds, softball-size hail, and tornadoes.” You notice that, yes, moist air from the Gulf had been entering the state lately, resulting in a stretch of sultry, humid weather.

Sometimes the Gulf humidity becomes tornado fuel because of a cold front being pulled to our latitudes from up north by the meanderings of the jet stream. On that weekend before May 24 a dry line approached Oklahoma from the west, coming from the deserts to the southwest and over the lee side of the Rockies. The moist and dry air masses soon to be colliding over Oklahoma could result in vigorous convection, and the wind currents aloft were showing vertical shear, making the outlook ominous. On Sunday and Monday the television station meteorologists warned emphatically of “high probability of severe weather on Tuesday.”

As Tuesday begins, you batten down the hatches. You shut the horses in the barn and latch all the doors and windows. You

put your dogs in their kennel inside the house. As you leave for work, you glance at your house and hope for the best as you drive away in the high-mileage car with the faded paint. You leave the “good car” in the garage, thinking about that hail. Here we go again, you think, another May day with the familiar threat of severe weather. Somebody is going to get it, but Oklahoma is big and tornadoes are small. The sirens will probably sound. It promises to be quite a show. Which neighbors with storm shelters are at home—just in case . . . ?

That morning the surface wind comes out of the south at 20 mph, carrying warm humid Gulf air with it. As the dry air mass moves into the state from the west, it shoulders its way beneath the less dense moist air. Illustrating vector addition, thunderstorms in Tornado Alley typically travel from southwest to northeast.

This morning, behind the scenes at the television stations and at the Storm Prediction Center in Norman, meteorologists are looking at their balloon sounding data and measuring “CAPE,” the convective available potential energy. The numbers from their instruments spell trouble for late afternoon and into the evening.

As the day moves along initially under a bright blue sky, let’s follow a “parcel” or bubble of air that starts out as part of the moist air mass brought up from the Gulf. As the day progresses the dry line moves across the state and undercuts these bubbles of moist air, lifting them to higher altitudes. If the conditions are primed for severe weather, eventually they are lifted to elevations where the bubble finds itself hotter than the surrounding air. Now the air bubble becomes buoyant and accelerates upward. A roaring convection, interacting with wind shear aloft, could bring to pass the anticipated trouble.

In summary so far, the “standard model of supercells,” if there is one, includes three conditions:

1. A strong lifting mechanism, such as an invasive dry line or cold front, boosting humid air to altitude;
2. Condensation of the humidity at altitude to release latent heat and drive accelerated buoyancy with a vigorous updraft; and
3. Vertical shear.

In the remainder of this article I will go over these points in more detail, describing how they work together to spawn tornadoes. However, before going there we must set the stage. Normal weather patterns, dominated at the continental scale by horizontal flows, offer the unperturbed system on which local convection and vertical shear are perturbations.

Disclaimer: I am a physicist, not a meteorologist. But in this study my appreciation of meteorology and meteorologists has been greatly enhanced. The level of accuracy I can offer here is comparable to that of the metropolitan traffic physicist who discovers that traffic flows from the suburbs to downtown in the morning and reverses in the evening. That’s not wrong, but of course there’s much more to the story. So simplified a picture provides a useful place to start, however. If I can offer in this article a similar simplicity regarding supercells, mesocyclones, and tornadoes, I will count its purpose as being served. Let us begin by discussing unperturbed air flow patterns in the atmosphere.

The Unperturbed Atmosphere

Air consists of a mixture of gases dominated by molecular nitrogen and oxygen, about 78% and 21% by volume, respectively, in dry air. Carbon dioxide, hydrogen, methane, ozone, the noble gases, and pollution make up the balance, along with water vapor where the air is humid. Humid air is less dense than dry air (other parameters being equal) because the water molecule (H_2O) weighs less than nitrogen (N_2) and oxygen (O_2).

Here on the surface we live at the bottom of this ocean of air. We feel the weight of all the air overhead. That weight per unit area, the atmospheric pressure, decreases with elevation.

The flux of solar power received at Earth’s orbit is about 1400 W/m^2 . Some 40% of it gets reflected directly back into space. The atmosphere is relatively transparent to radio and visible wavelengths, which warm the oceans and land masses. They reradiate some of their internal energy, mostly in the infrared wavelengths, for which water vapor and carbon dioxide are especially good absorbers, thereby heating the atmosphere. When its temperature increases, a parcel of the atmosphere expands. The pressure, density, and temperature of a parcel of air are related by an equation of state, the simplest model being that of an ideal gas.

Above different regions of the landscape, the air masses carry different values of temperature and pressure. These state variables often show sharp discontinuities, called “fronts,” between regions. When cold air advances into warmer air, we speak of a “cold front”; when warm air advances into cooler air, we describe it as a “warm front.” The leading edge of a colder air mass, the cold front, is denoted as in Fig. 4a and a warm front as in Fig. 4b. The warm air slides over the cold air due to the density difference (Fig. 4c).

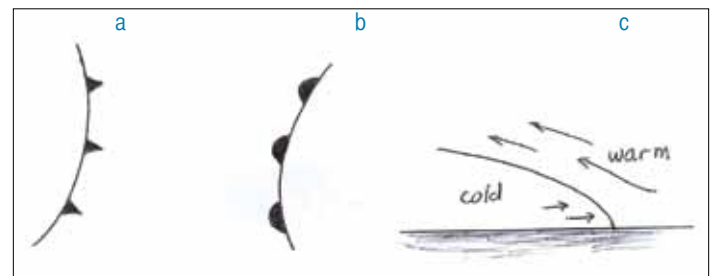


Fig. 4. (a) Notation for a cold front and (b) a warm front. (c) Side views of a cold front colliding with warm air.

On the surface where a mass of dry air meets a mass of moist air, one speaks of their boundary as the “dry line.” A dry line is similar to but not synonymous with a “front.” A front advances but does not recede, whereas a dry line, though it advances during the day, might recede at night when the winds die down and the dry air entrains moist air from the other side of the line.

The air generally becomes cooler and drier with height, a trend that continues up to about 11 km at the poles and some 17 km above the equator. This region of the atmosphere defines the “troposphere,” where $dT/dz < 0$, where T is temperature and z denotes the vertical distance above sea level. Meteorologists call $-dT/dz$ the “lapse rate”; the troposphere carries a positive lapse rate.

Above the troposphere the temperature *increases* with height (negative lapse rate, an “inversion”), a layer called the “stratosphere,” which starts at around 50,000 ft. This counterintuitive temperature gradient occurs because in the thin air at these high altitudes ozone is increasingly exposed to ultraviolet light, breaking it down from O_3 to $O + O_2$. The O and O_2 then rejoin in an exothermic reaction.

Air is a poor thermal conductor and thus, with its low but nonzero viscosity, convection serves as its chief mechanism for heat transfer. Due to vertical convective mixing, weather action occurs in the troposphere. The temperature inversion in the stratosphere discourages cooler troposphere air from rising into it, so the stratosphere carries little convective mixing and is relatively stable. The region of transition from the troposphere to the stratosphere is called the “tropopause.”

Within the troposphere large volumes of air expand and contract due to differential heating, and the resulting pressure gradients drive winds which, along with Earth’s rotation to provide extra steering, push the atmosphere to and fro. As with swells in the water ocean, even over flat ground the depth of this air ocean varies, with “ridges” and “depressions” whose pressure “highs” and “lows” are felt at the surface beneath the column of air overhead. Meteorologists find it convenient to map vertical air dynamics, not in units of altitude but in units of pressure, with “isobaric surfaces.” An isobaric surface is a locus of points having the same pressure. In meteorologist talk, one speaks of, say, “the 500 mb surface.” The “mb” stands for millibar. One bar, the unit used by meteorologists since 1909, approximately equals the atmospheric pressure at sea level.[4] Near the surface the pressure decreases with altitude at about the rate of 100 mb/km, so that we encounter 500 mb about 5 km aloft. But as we go up the air density also decreases, so the relation between pressure and altitude is logarithmic rather than linear. For example, 300 mb occurs at about 9 km above sea level. Under a pressure “high” the air gets more compressed; instead of the surface carrying 1000 mb it may carry a bit more, which means the 1000 mb surface lies at a higher altitude than usual (Fig. 5). Under “lows” the 1000 mb surface sags downward, as do the other isobaric surfaces above it.

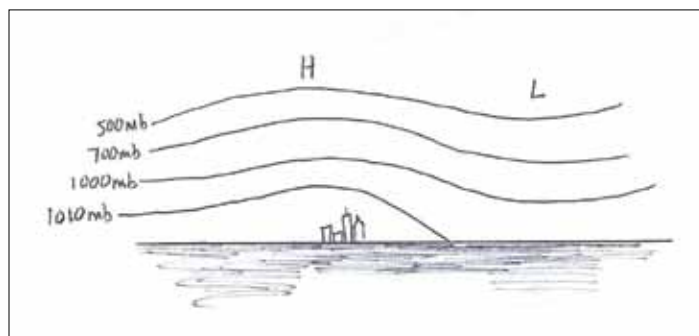


Fig. 5. Isobaric surfaces.

Pressure gradients drive the winds. Let us examine the pressure and velocity fields over our heads. Adjacent layers of moving fluid tug on one another through viscous friction, setting up a velocity gradient as illustrated in Fig. 6.

On the ground itself the air velocity equals zero, but

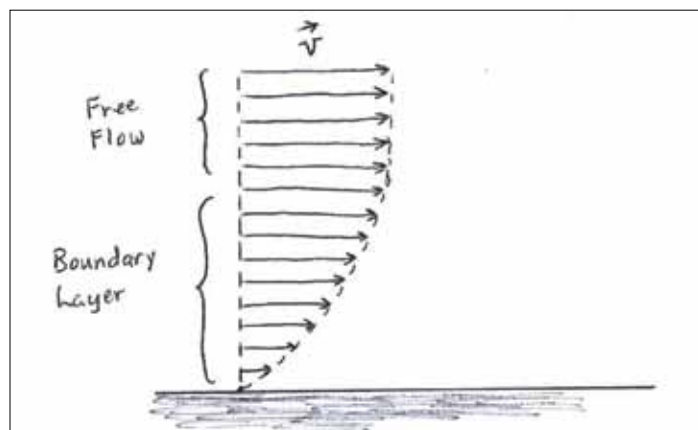


Fig. 6. Shear due to viscosity.

the speed increases away from the surface and into the flow. Sufficiently far from the surface, where the layers of fluid no longer slide over one another, the velocity field is said to be that of “free flow.” The region of nonzero velocity gradient between the surface and the onset of free flow forms the “boundary layer.” Above the landscape the wind’s boundary layer is on the order of a kilometer thick. Wherever the boundary layer separates from the surface due to surface topography, vegetation, buildings, clouds, and so forth, laminar streamline flow breaks up into turbulence and eddies.[5] Such mixing also occurs at a front between colliding air masses, when opposing velocities with viscosity whip up eddies or whirlpools. In the free-flow regions there is less turbulence and mixing, so on short time scales the winds are steadier and the temperatures more stratified on high than at the surface.

The Earth’s atmosphere is very shallow compared to the planet’s radius. The relative dimensions can be visualized by stretching tissue paper over a desktop globe. Therefore large-scale atmospheric dynamics are dominated in the first approximation by the horizontal components of pressure gradients and velocities. Let us consider them first.

Figure 7a illustrates a region of excess pressure, a “high,” and another region of diminished pressure, a “low,” in a sheet of atmosphere. The pressure gradient drives wind from the high to the low. If the Earth’s rotation could be turned off, the winds would diverge from the high and converge into the low as directly as possible, analogous to an electrostatic dipole field.

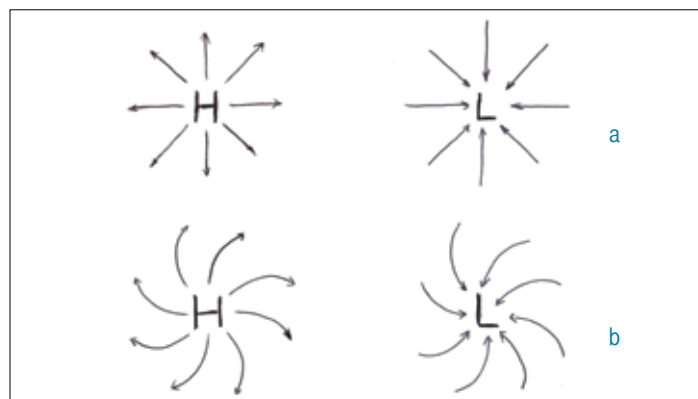


Fig. 7. Pressure gradients driving the velocity field (a) without and (b) with the Earth’s rotation.

However, due to the Earth's rotation, observers aboard the reference frame on our planet's surface experience two "fictitious" forces, the Coriolis force and the centrifugal force (see Appendix, www.spsobserver.org/tornadoes). The centrifugal force is usually negligible for continental-scale air movements. The Coriolis force, in contrast, plays a large role. Analogous to a magnetic force, it deflects the velocity vector of moving air in a direction perpendicular to \mathbf{v} . Because of the Coriolis force, the velocity fields of Fig. 7a are modified into those of Fig. 7b, where (in the Northern Hemisphere) the moving air circulates counterclockwise about a low (a "cyclone") and clockwise about a high (an "anticyclone").

Like magnetic force, the Coriolis force changes only the *direction* of a particle's velocity. This velocity continues to be deflected unless the Coriolis force gets cancelled by another force. A horizontal pressure gradient and the Coriolis force cancel when the wind velocity is tangent to an isobar, a situation called "geostrophic flow." The jet streams are an example of a geostrophic flow.

Now let us bring in the vertical dimension. In static atmosphere the weight of a horizontally oriented slice of air is cancelled by the vertical pressure gradient. But air is seldom static. Important vertical velocities come from mechanical lift and convection. Common forms of mechanical lift occur when an air mass glides up the slope of a mountain range; gets a push from low-altitude thermals; or, as commonly seen in the prairie, when two air masses of different densities collide and the warmer, moist, less dense air slides up over the cooler, drier, denser air mass (e.g., at a cold front or dry line). Whatever the lifting mechanism, the lifted air rises too quickly for conduction to be an effective energy-transfer mechanism, so as the pressure decreases the air parcel expands and cools adiabatically. If its density becomes less than that of the environment, then it becomes buoyant and moves upward. Any mechanism that warms the atmosphere at low altitude and cools it at high altitude promotes convection.

On a global scale, between the equator and the poles the rising of warmer air and the sinking of cooler air, along with horizontal flows aloft and at the surface, complete the circuit for the large-scale circulations responsible for trade winds and for the great deserts such as the Sahara in Africa, the Gobi in Asia, and the Sonora of the southwest United States and northern Mexico. Because equatorial terrestrial surfaces receive a higher flux of solar energy than polar latitudes, equatorial lands and oceans

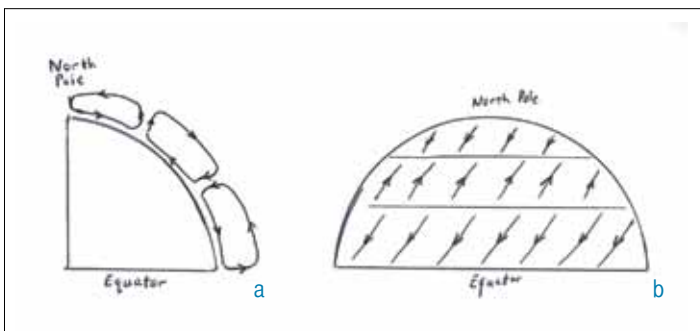


Fig. 8. (a) Surface winds and (b) the convection cells responsible for them in the Northern Hemisphere.

are warmer than that farther north or south. The heated air near the equator drops in density and rises in elevation. As the air rises it expands and cools, hitting the temperature inversion at the stratosphere. In the northern hemisphere, some of this moving air flows toward the north, from whence the Coriolis force gives the velocity vector an eastward component. At around 30° latitude this air sinks down and some of it splashes to the south, where the Coriolis force gives these surface winds a westward component, setting up the sailing ship's trade winds (Fig. 8a). Great deserts lie at these latitudes because their landscapes sit beneath the dry downdraft side of these large circulation cells.

Cold air near the poles sinks and creeps across the surface toward equatorial regions, and in so doing the Coriolis force gives its velocity a westward component. When it reaches about 60° latitude it has warmed sufficiently to rise by convection. The polar and equatorial convection cells drive a third cell between them (Fig. 8b) whose surface winds produce trade winds from southwest to northeast.

At the boundary between the cells high in the atmosphere, pressure isobars there set up distinctive geostrophic flows—the jet streams. These rivers of fast-moving air are typically about 100 miles wide and 3 miles thick. The subtropical jet stream flows about 10–16 km high, and the faster polar jet stream around 7–12 km aloft. Their speeds range from 50 to 250 mph. In both hemispheres all four main jet streams move west to east, with wavelike meanderings north and south. Thus in February in Oklahoma, one week can be great motorcycle riding weather while the next the jet stream dips south out of Canada to visit upon us a hard freeze that makes us empathize with our friends in Michigan's Upper Peninsula. Thanks to Bernoulli's equation [Eq. (A7) in the Appendix], higher speed means lower pressure. Thus a jet stream passing overhead represents a high-altitude river of relatively low pressure, which helps pull convection aloft in storms trying to form beneath them.

The "standard model" (if it exists) of tornado formation requires an initial lifting mechanism and vigorous upward convection. The third ingredient is "vertical shear," where the horizontal component of the wind velocity increases in magnitude and changes direction with increasing height. Together, convection and vertical shear can brew up thunderstorms packing considerable angular momentum, as we shall see. We will return to vertical shear later. The thermodynamics for convection is studied by meteorologists with their "skew- T -log- P diagram," or "skew- T " for short.

The next issue of *The SPS Observer* will feature Part II of "Tornado Physics," highlighting the role of thermodynamics and vertical shear in tornado formation. Part II also describes the author's inspiration for this story—a much too close encounter with an F5 in May 2011.



The Physics of Tornadoes: Part 2

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This article is available in its entirety at www.spsobserver.org

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—Chief Luther Standing Bear (1929)

Skew-T Diagrams

To prepare the way for thermodynamics in thunderstorms, let us recall our old friend the Carnot engine. The Carnot cycle was invented by Sadi Carnot in 1824 to calculate the maximum efficiency obtainable, in principle, from an engine that operates between two temperatures. Therefore the Carnot cycle features two isothermal and two adiabatic processes. It can be represented in pressure-volume P - V space as shown in the familiar diagram of Fig. 9a and in the temperature-entropy T - S space as in Fig. 9b. The area enclosed by the P - V diagram equals the net work done by the engine in one cycle. The area enclosed by the T - S diagram equals the net heat put into the engine per cycle. By the conservation of energy, these work and heat values per cycle are equal.

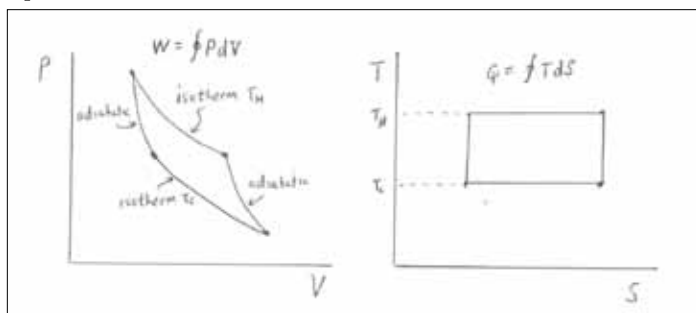


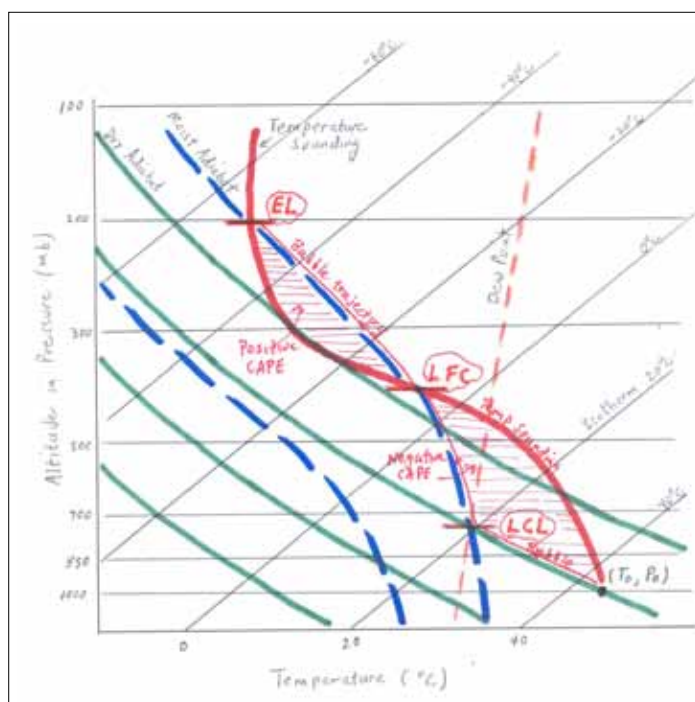
Fig. 9. The two adiabats and two isotherms of the Carnot cycle in (a) P - V space and (b) T - S space. The areas enclosed are the net work per cycle done by the engine and the net heat per cycle put into the engine, respectively.

As a bubble of air gets lifted, its pressure and temperature decrease adiabatically. Recall that for an adiabatic process with an ideal gas, $PV^\gamma = \text{const.}$, where γ (~ 1.4) denotes the ratio of specific heat at constant pressure to that at constant volume. This adiabatic expression can also be written $T \sim P^{(\gamma-1)/\gamma}$. Therefore on a graph of P vs T the adiabatic processes make declining curves. Notice that the slopes of the adiabatic curves correspond to dz/dT , the inverse of the lapse rate. These curves that track the adiabatic ascension of an air parcel are called “adiabats.” As adiabatic changes of state, transformations along these lines proceed at constant entropy.

When plotting P on the vertical axis, the spacing of equal pressure increments grows logarithmically with altitude. Instead of orienting lines of constant temperature vertically above the linear T axis, meteorologists slant the isotherms at 45 degrees,

forming the skew- T graph (Fig. 10), making the meteorologist’s version of a pressure-temperature graph for tracking thermodynamic states in the atmosphere.

Fig. 10. A skew- T diagram.



Since the atmosphere is *not* in thermodynamic equilibrium during interesting weather, you would be justified in asking how we can speak of “the state” of the atmosphere in the first place. Your objection would be correct: the atmosphere as a whole does not have a state. However, we will consider a “parcel” of air, a “bubble” of atmosphere small enough so we can speak of its temperature and pressure state as it moves vertically through the ambient atmosphere whose temperature and pressure vary with elevation.

The adiabats depend on the value of γ , and thus on the relative humidity of the parcels of the air. Two sets of adiabats are shown in Fig. 10: the dry adiabatic lapse rate (solid curves) and the moist adiabatic lapse rate (dashed curves) for air saturated with moisture. These rates are about $9.8^\circ\text{C}/\text{km}$ for dry adiabats and $4.1^\circ\text{C}/\text{km}$ for moist ones (the slopes change with increasing pressure on the skew- T diagram because of the logarithmic spacing of the vertical axis). When pressure

vs temperature data for today's atmosphere is plotted on the skew- T diagram, the departures of the data's trajectory from the gracefully curved adiabats allows meteorologists to predict the behavior of a bubble of air that (by whatever mechanism) finds itself set into vertical motion. Just as the area in the T - S state space represents heat available to do work, likewise on the skew- T diagram the area between two given adiabats and two given isotherms represents such energy.

To routinely gather this temperature vs pressure data, the National Weather Service, with corresponding agencies from over 90 countries, twice daily launch over 1000 balloons. About 70 launch sites exist in the continental United States. Each balloon carries a shoebox-size payload of instruments and transmitters, a "radiosonde." The balloons reach altitudes in excess of 30 km (18 miles \approx 100,000 ft). The temperature vs pressure data immediately gets plotted on the skew- T diagram, a trajectory called an "environmental temperature sounding." This sounding shows how the atmosphere's actual temperature varies in a column of air that towers over the heads of the people in, say, Oklahoma City this morning.

Let us consider what happens to a bubble of dry air that gets lifted up, say, by a cold front sliding beneath it. The bubble starts out near the surface around the 1000 mb level at some temperature T_0 and pressure P_0 . As it lifts, the bubble decompresses and cools. We track its change of state along the dry adiabatic trajectory that happens to pass through the point (T_0, P_0) on the skew- T diagram. The adiabatic trajectory cuts upward across isothermals, heading in the direction of decreasing temperature.

If this morning's environmental sounding happens to coincide with the dry adiabat followed by our bubble, then the bubble and the surrounding air through which it passes always have the same temperature, giving the bubble neutral buoyancy. In that case, if the bubble were lifted to an elevation and then released, it would just sit there. The bubble and its environment are in thermodynamic equilibrium; there will be no severe weather on this day.

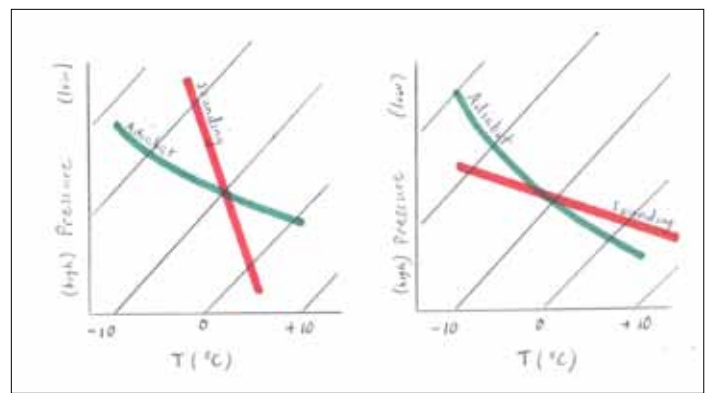
What happens if, at a given pressure (altitude), the temperature sounding does not coincide with the bubble's adiabat? Suppose that at some elevation the sounding line slope is steeper than the parcel's dry adiabat (Fig. 11a). At the point where the adiabat and sounding line intersect, the bubble and the surrounding air are at the same pressure and temperature. But as the bubble rises further its state on the diagram moves to the left of the sounding line, leaving the bubble cooler and thus denser than the surrounding air at the same altitude. The bubble will stop rising and start sinking. If it picks up sufficient speed while sinking to overshoot the static equilibrium point, after the overshoot the bubble's temperature coordinate lies to the right of the sounding line at the same pressure. Now the bubble finds itself warmer than the air around it so it pops back up. The air bubble's vertical motion undergoes restoring force oscillations and no thunderstorms occur.

From elementary mechanics we recall that changing the sign on the spring constant of a simple harmonic oscillator turns its motion from oscillations into a runaway exponential. Something

similar happens on the skew- T diagram if we change which line has the steeper slope.

Recall that where the adiabat and sounding line intersect, the bubble and the surrounding air are at the same pressure and temperature. But when the adiabat has the steeper slope, as the bubble gets lifted past the equilibrium position its state on the diagram moves to the right of the sounding line, which means the bubble is warmer than the ambient air around it at the same pressure. Now less dense than its surroundings, the bubble becomes buoyant and *spontaneously* rises (Fig. 11b). Conversely, if the bubble could somehow be pushed down from the state of intersecting slopes, it would be driven further downward. Under these circumstances the atmosphere in this region of (T, P) space is unstable against vertical motion.

Fig. 11. (a) Stability and (b) instability as revealed on the skew- T diagram.



So far we have followed a bubble whose state tracks a dry-air adiabat. But tornadoes come out of thunderstorms and thunderstorms imply rain—pelting rain with wind and lightning and thunder. Tornadoes develop in humid air for a reason: the humidity, through the latent heat it releases when condensing, provides the high-octane fuel to drive a bubble's instability toward a violent end. Let us now consider lifting a bubble that carries humid air.

If the bubble on the ground is not already a fog, its water is carried in the gaseous phase. As it gets lifted it starts out by following the dry adiabat curve passing through its initial (T_0, P_0) coordinates, but it carries less density than dry air. So long as the lifting continues, eventually the bubble reaches the altitude where its temperature drops to the "dew point."

The dew point, by definition, is the temperature to which humid air must be cooled at constant pressure for the water to precipitate out. We see dew on the grass some mornings because overnight the temperature dropped to the dew-point value for the ground-level pressure. Of course, as our bubble of air rises its pressure changes with altitude. However, the dew point rescales from its ground-level value by a proportionality factor m called the "mixing ratio," the ratio of the density of water to that of dry air. Since the mass of a given bubble stays the same as it moves vertically, the mixing ratio is independent of z . We can therefore write the variation of dew point with altitude as $T_{\text{dew}}(z) = mT_{\text{dew}}(0)$ and plot $T_{\text{dew}}(z)$ on the skew- T diagram, the mixing-ratio line. It will cut across the isotherms, from warmer to

colder. Air that can hold water in the vapor phase at ground level will become saturated aloft. The moist adiabat slopes are lapse rates for saturated air.

The radiosonde balloons also collect data on the air's humidity as a function of height. Translated into the dew-point temperatures, this data can also be plotted on the skew- T diagram (the dashed line in Fig. 11) to make the "dew-point sounding." This sounding shows the humidity that is already aloft, which can be pushed up with or entrained into the bubble as it rises.

Let us continue following our bubble of humid air being lifted by a cold front. The bubble started its trip aloft with water vapor mixed with the air but not yet saturated, and thus began its skew- T trajectory on the dry adiabat passing through (T_o, P_o) . But when this trajectory crosses the mixing-ratio line, the bubble becomes saturated and the water vapor in it precipitates out, turning our bubble into a cloud. The altitude where this occurs is called the "lifted condensation level" (LCL). It accounts for the flat bases of clouds.

This phase change of the water from vapor to liquid releases an enormous amount of latent heat into the bubble, some 2.26 million joules of energy for each kilogram of water.[6] After condensation, if the foggy bubble continues to be carried aloft, it continues to expand and cool at the moist adiabatic lapse rate, which is less than the dry rate thanks to the shot of latent heat.

What happens next depends on the relative slopes of the environmental sounding and the moist adiabat at the LCL event; or, to say it another way, it depends on whether the bubble's adiabatic state lies to the left or to the right of the temperature sounding at LCL pressure. If the LCL event happens on the left of the temperature sounding, then the newly formed cloud is cooler and denser than the air around it and it sinks back down. The droplets in the parcel dry out and the show is over. We often see this in fair-weather cumulus clouds that grow and dissipate throughout the day.

If clouds form with insufficient latent heat to warm the bubble to accelerated buoyancy but carry enough moisture to prevent dry-air entrainment, then the cloud may persist and we walk around under partly cloudy or overcast skies. Should sufficient water reach the LCL so the droplets coalesce, we are treated to a refreshing shower.

If enough bubbles laden with humidity reach the LCL so that many clouds appear but the sounding temperature above them increases (a temperature inversion), the storm chasers with their electronics-stuffed vans start getting frustrated. The clouds accumulating over their heads cannot get into the cooler regions at higher altitude. The storm chasers start muttering about "the cap" overhead.

During the warm months many days begin with a cap because the ground cools overnight, and with it so does the air immediately above the ground. When the air higher up stays warmer throughout the night, the day begins with a temperature inversion, the cap. Throughout the day as the spring or summer sun beats down, the land warms and heats the air above it so that by late afternoon the low-altitude air has become warmer

than the air higher aloft, dissolving the cap from below. That is why most severe weather starts firing up toward late afternoon. But as long as the cap stays put, the clouds that congregate above the LCL cannot go much higher.

The probability of turning this situation into severe weather is greatly enhanced if the lift from an invading cold front or dry line is strong enough to continue lifting the bubble past the LCL, up to altitudes where the moist adiabat on the skew- T graph intersects the temperature sounding line. There the bubble and the surrounding air have the same temperature. The altitude where this occurs is called the "level of free convection" (LFC). If the bubble has not yet rained out its moisture, any additional upward lift continues pushing it along the moist adiabat to states that lie to the right of the temperature sounding line. Physically, the bubble is now in a region where it is warmer than the ambient air. The bubble becomes vigorously buoyant and no longer needs the dry line or cool front to get a lift. Like a beach ball released from the bottom of a swimming pool, the bubble accelerates upward. Air from below rushes upward to take its place, at speeds comparable to automotive boulevard traffic. A vigorous convection has started. The thunderstorm has started to rumble. We see the cumulus clouds boiling upward into great mushroom clouds and exploding to towering heights in the upper troposphere. Lightning and thunder start their spectacular fireworks shows. Down below, the storm chasers run for their vans and switch on their radios.

Up above, the rising saturated bubble continues to accelerate upward, expanding and cooling until it reaches the height where its temperature drops to equal that of the ambient air, the "equilibrium level" (EL) where, as represented on the skew- T diagram, the moist adiabat once again intersects the temperature sounding line. Here the convection stops and the anvil forms. If the bubble is large enough and rises fast enough, it may carry sufficient vertical momentum to overshoot the EL, perhaps breaking into the stratosphere. There the ozone-induced temperature inversion inevitably stops the bubble's furious rise and the cloud spreads out into an anvil at some 50,000 ft. altitude.

Meanwhile, the vigorous convection has made a chimney flue for other warm moist air beneath it to follow. If this enthusiastic convection is accompanied by an equally vigorous vertical shear, the thunderstorm may organize into a supercell.

When the morning temperature sounding was plotted, the meteorologists estimated the value expected for the afternoon's highest temperature near the surface and traced the projected trajectory on the skew- T diagram from which a bubble would set off. From the mixing-ratio line and the dew-point sounding the meteorologists estimated the humidity carried by the bubble and the moisture that might be entrained by it on the journey aloft, projecting the most likely LCL, LFC, and EL. An important number that meteorologists derive from these estimates is the energy per unit mass likely to be carried by the convective air rushing aloft, the energy that provides thunderstorms with fuel. On the skew- T graph this energy equals the area between the environmental sounding and the bubble's adiabat. Below the LFC

this energy is negative, because here the bubble needs an external boost to be lifted up, but above the LFC and below the EL, the bubble is buoyant and its energy positive. This energy density available to drive convection is called the convective available potential energy, or “CAPE.” A positive CAPE up to about 1000 J/kg describes “moderately unstable” conditions; a CAPE above 3500 J/kg means that conditions are poised for severe weather.[7]

On May 3, 1999, a tornado outbreak occurred in central Oklahoma that included a mile-wide, long-tracking F5 tornado packing vortex speeds of over 300 mph. This tornado plowed through densely populated neighborhoods in Moore, Del City, Midwest City, and other communities in the Oklahoma City region, resulting in the most destructive single tornado in recorded history so far. The supercell that spawned it that afternoon began with a CAPE of 5885 J/kg.[8] A large CAPE does not guarantee gargantuan tornadoes, but it certainly provides sufficient energy for them.

Prolific breeders of tornadoes are “supercells,” a term introduced by British meteorologist Keith Browning in 1964. A supercell can be some 20 miles wide and dominate the weather locally for several hours. Besides carrying high unusually high CAPE, they also differ from ordinary thunderstorms in longevity and size thanks to vertical wind shear, which we examine next.

Vertical Shear

For a thunderstorm to develop into a supercell, strong horizontal winds with vertical shear at high altitudes are necessary. Vertical shear means that with increasing height the horizontal components of \mathbf{v} increase. If the storm exhibits both large CAPE *and* vertical shear, then as the cumulus cloud grows higher the horizontal winds in its upper reaches get stronger, which makes the towering cumulus cloud lean over. Thus its rain and hail will not fall straight down through the rising column of convection, nor will its gust front impede the inflow of warm moist air at the base. The storm continues to grow and grow and grow. . . . The television meteorologists start interrupting programming to talk excitedly about the storm “building.”

The wind velocities, and thus vertical shear, are also measured by the radiosondes hanging from their balloons. Consider a reference frame having its origin, say, in your front yard, with the x axis pointed east, the y axis going north, and the z axis toward the zenith. By definition the “vertical shear” \mathbf{S} takes the form in this frame’s coordinates

$$\mathbf{S} = \frac{\partial v_x}{\partial z} \mathbf{i} + \frac{\partial v_y}{\partial z} \mathbf{j}. \quad (1)$$

An example of such a velocity field is illustrated in Fig. 12.

Any shear (vertical or otherwise) can produce “vorticity,” a nonzero curl of the velocity field, $\nabla \times \mathbf{v}$. For instance, in the

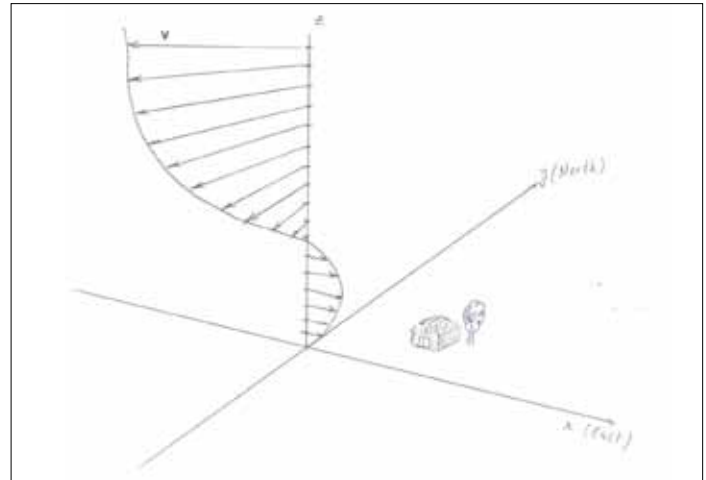


Fig. 12. Vertical shear, with “veering winds.”

example of Eq. (1) we find that $\nabla \times \mathbf{v} = S\mathbf{j}$. The right-hand rule for this cross product means that as seen looking toward the north, the vorticity is horizontal and spins clockwise. Such vortices happen in many wind-shear situations, but due to fluctuations and turbulence they typically come and go, like the leaves whirling around momentarily above the bed of your moving pickup truck.

If the vertical convection rises through a persistent large horizontal vortex (for example, a roller caused by winds swooping eastward over the Rockies), the horizontal tube of circulation gets lifted and bent into the shape of an upside-down U, one mechanism for generating a “mesocyclone.” Suppose the horizontal vortex tube was originally aligned north–south. When bent into an inverted U, the section south of the convection spins counterclockwise (viewed from above), the “cyclone,” and the section to the north spins clockwise, the “anticyclone” (Fig. 13).

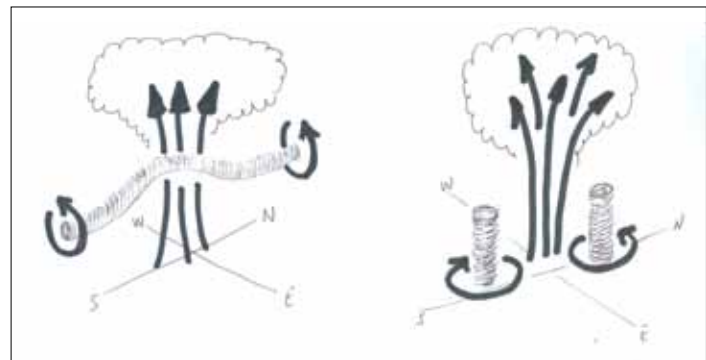


Fig. 13. One mode of vertical vortex formation.

These large masses of rotating air, kilometers in diameter, call for watchfulness but are not yet tornadoes. Meanwhile, the original bubble continues to expand as it rises atop its convective chimney, pushing the cyclone and anticyclone apart.

On the upshear (tail of the \mathbf{S} vector) side of the bubble, the pressure is higher than at the same elevation on the downshear side. This is so because at any point the velocity vector is the

superposition of the vertical and horizontal winds (Fig. 14). Thus the net velocity vector has a larger magnitude and thus produces lower pressure on the downshear side. This horizontal pressure gradient, caused by these cross-bubble pressures, steers the bubble horizontally as it continues to rise.

What happens next depends on the change in direction of \mathbf{S} with increasing height. Should \mathbf{S} turn clockwise with increasing altitude, one speaks of “veering winds.” The contrary cases of counterclockwise rotation for \mathbf{S} are called “backing winds.” Veering winds occur far more frequently in the Northern Hemisphere than do backing winds. Rising through veering winds, the bubble pushes the adjacent cyclone before it, deflecting it south. With the anticyclone left behind to dissipate, we now have a large counterclockwise-circulation mesocyclone. Things are getting interesting.

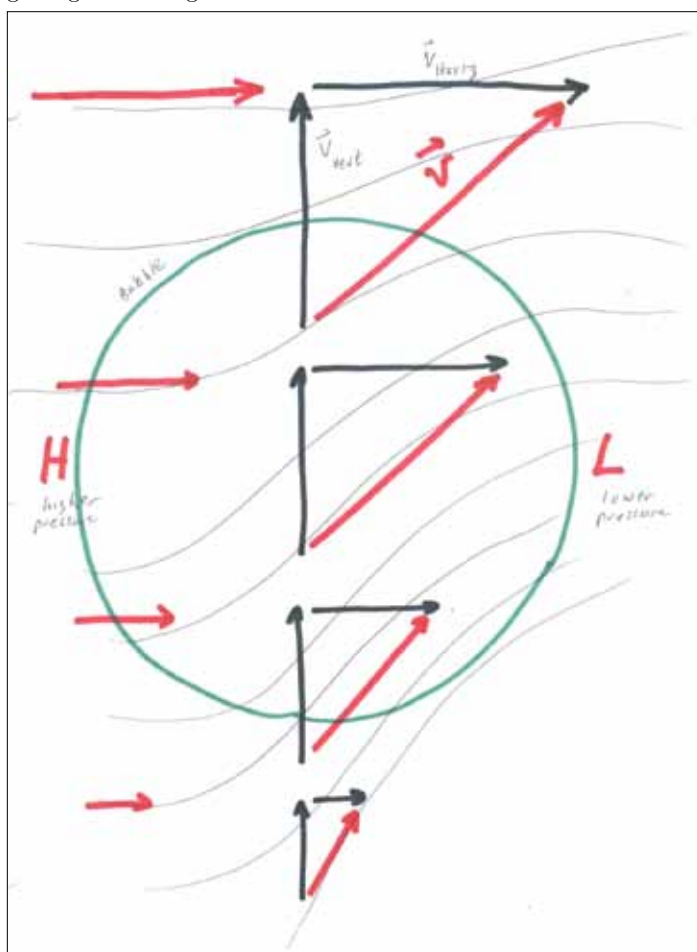


Fig. 14. Shear pushing a bubble horizontally.

The meteorologists at the television stations preempt afternoon game-show programming to announce something serious: the appearance of a “circulation.” They name the towns in its path, giving for each one the estimated time of the circulation’s arrival. In the field the storm chasers jockey for position, each striving to be the first to spot the “wall cloud”; brightly painted helicopters circle the storm and beam telephoto images to an attentive audience below; TV screens switch back

and forth from color-coded Doppler images superimposed on street maps to cell phone chaser images photographed through wiper-swished windshields and described over scratchy audio; all with breathless nonstop chatter and TV station name-dropping. It is all very exciting and makes for great entertainment. Across the state, in family rooms, sports bars, and coffee shops, all eyes are glued to television screens.

The TV meteorologists seem to enjoy the storms. No doubt they see the beauty, even though they know that somebody somewhere is going to get flattened. Sometimes they seem to hype the storms to excess; you listen to the TV and doom seems about to drop onto your head. Then you go outside and look at the real sky with your own eyes. Most of the time you see the towering cumulus clouds off in the distance, illuminated from within by spectacular nonstop lightning while their mushroom cloud summits catch the last pink rays of sunset. From a distance it *is* beautiful, but these storms are genuinely dangerous to those under them (lightning kills more people than tornadoes), and the meteorologists provide a genuine public service, even while their TV stations compete for ratings and try to outdo each other with gee-whiz technology, amazing images, and catchy slogans (“Keeping you Forewarned,” “Eye on the Storm,” etc.).

It has often been said that when storm sirens start blaring their mournful air raid wails, you can tell who comes from Oklahoma: they are the people who run outside to look while everyone else scurries for the shelter. There is some truth in the caricature. But it must also be said that over the years the TV meteorologist’s warnings have been responsible for saving untold numbers of lives. We may run outside to take a photo, but when it gets dangerously serious in our neighborhood we dive for cover, even if it’s at the last minute.

To “tornado” the mesocyclone vortex has to tighten up to smaller radius and higher speeds, and extend all the way from the clouds to the ground. In some cases the tornado grows downward; other mechanisms feed low-level horizontal vortices into the mesocyclone. Sometimes both happen. One downward-growing method, the “dynamic pipe effect,” proceeds as follows (Fig. 15). Pressure within the mesocyclone is low, so air feeds into the growing maelstrom from below, increasing in speed as it spirals in and up. As the inflow speeds up its pressure drops. In this way a region of circulation and low pressure propagates downward, soon to appear as a “wall cloud,” a cylinder of rotating cloud that sticks out prominently beneath the main cloud base.

As more air rushes in the angular velocity of the vortex increases, and by the conservation of angular momentum (approximate, due to friction) the diameter of the vortex decreases. When the vortex connects the cloud and the ground, it has become a tornado. The sirens shriek and moan, and the television meteorologists say—with earnest seriousness—“If you live in Piedmont or Cashion, go to a shelter NOW. If you don’t have a shelter, go to the smallest room in the center of your house away from windows and pull a mattress over your head. But TAKE COVER NOW!”

The tornado might have a constant radius from cloud to ground (a “stovepipe” tornado), or its radius may increase with

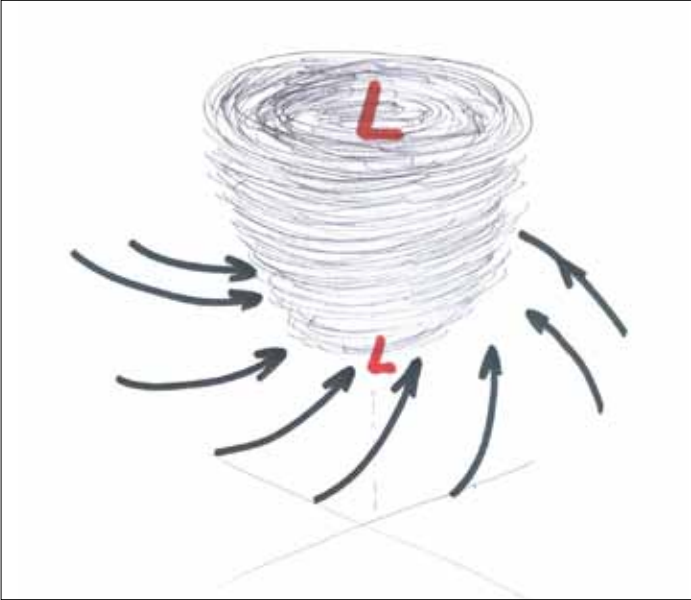


Fig. 15. Feeding the vortex.

height (a “wedge” tornado), or it may narrow down to a ropelike vortex, which often happens as the twister dissipates. But the thin rope tornadoes still pack wicked speeds. The speeds in the tornado vortex can range from ~70 mph (F1) to ~300 mph (F5) at the ground level and increase with height. Some large mesocyclones will spawn multiple tornadoes, or a large vortex may sprout satellite vortices orbiting about it, like planets spinning on their axes while orbiting their star.

The tornado continues moving across the landscape as long as the ambient horizontal winds push the supercell along. The vortex continues grinding away until its inflow gets cut off. For instance, when the heavy rain begins to fall, the general circulation wraps the mesocyclone and its tornado in a revolving curtain of downpour. As the tornado becomes rain-wrapped, the inflow of warm moist air is slowly quenched and eventually the tornado dissipates.

Close Encounter

Although my immediate family has lived in central Oklahoma only 25 years, well short of the millennial mean-free time between consecutive tornado visitations to a given site in this region, our neighborhood was honored with an impersonal visit by the F5 tornado of May 24, 2011. Shortly before 5 pm my wife Rhonda and I were trying to drive home during the storm. Approaching from the south and listening to the action over the car radio, we could see only a vast blackness filling the space between the cloud deck and the ground. The blackness moved rapidly from the southwest to the northeast. Seeing that we were going to intercept it, we turned around, headed west, then turned north again, but were turned back twice by police roadblocks. Cell phone calls to our neighbors produced no response.

Detouring repeatedly through back roads, we finally came into our neighborhood from the northeast. Every tree, house, and outbuilding was gone.

Turning up your street into such a scene is rather disorienting (Fig. 16). One thinks, with deliberation: Despite the probabilities, *this is real*. As we jumped out of the car, our neighbors greeted us with relief; they had been looking for us. Then Rhonda ran for the horses and I ran for the dogs. I remembered that a colleague’s family lost their house in the epic tornado of May 3, 1999. Climbing into the rubble, I thought: They got through this, and we will too.

I hasten to add that we are doing fine, because Rhonda and I discovered a force abroad in the world far stronger than any mere F5 tornado: the abundant care and concern of so many kind and generous people. Before darkness fell and the police had to evict everyone from our neighborhood, some two dozen students, colleagues, friends, and family members were already there with us, helping us pick up the pieces, foreshadowing the caring concern to come. From them we saw that we would not have to face this alone. The morning after the tornado my son Charlie said to me, “Dad, you and Mom don’t have a house right now, but you are not homeless.” That was a perceptive distinction. Anne Frank was right: people really are good at heart.

Had we been in the house when the tornado struck, we would not have survived. Now we can still look forward to seeing our grandchildren grow up. Not everyone who began May 24 alive was so fortunate. Ten people lost their lives to tornadoes that day. The National Weather Service estimated that about 320 tornadoes occurred in May 2011, with 177 fatalities that month, 157 of those in the Joplin F5 tornado alone, two days before our tornado. The Joplin tornado was the deadliest May tornado since modern record keeping began in 1950, and it was the seventh deadliest in US history.[9] May 2011 has been so far the fourth-worst May on record for tornadoes.

On May 24 no human beings were killed on our street, although many four-leggeds lost their lives, including eight horses, among them our dear Annie, Rhonda’s equine companion for 11 years. Despite the losses, we are filled with appreciation. We have so much for which to be grateful.

Our twister was on the ground for 65 miles (Fig. 3), and when it passed through our neighborhood it was over half a mile wide and rain-wrapped, the blackness we saw earlier. When it hit our street, 17 of 21 houses were cleaned off down to their concrete slabs. The rest were so twisted that bulldozers later had to finish the demolition. A house, we learned, is a state of matter that exists between the Home Depot and landfill phases.

Tornadoes famously produce strange effects. You can’t find your roof or refrigerator, but you do find a toy car missing a wheel, and a few days later you find the wheel. The china cabinet gets smashed to smithereens, but a teapot that was inside survives intact. A bedroom gets swept aside into a pile of broken rubble, but a small night stand remains in its place like a lonely sentinel.

The vanishing of objects with large surface areas makes sense from Bernoulli’s equation: the kinetic energy density, proportional

Photo by Mark Winslow



Photo by Gary Unruh



Fig. 16. Our house, transformed into a high-entropy state, shortly after the tornado of May 24. I needed to clean out that garage anyway. I wish that high-speed cameras, like those used in the nuclear weapons tests of the 1950s, had been mounted throughout the house. I would like to see how our Silverado ended up on top of the Highlander. Photo by Mark Winslow.

to v^2 , gets turned into high pressure when the wind hits your piano. The wind's force equals this pressure integrated over the piano's surface area. With winds on the order of 300 mph, that force trumps mg for the roof, fence panels, doors, piano, sofa, appliances, bookcases, and dining room table. Carried aloft into yet stronger winds, they probably landed in pieces all over the next county; we never saw them again. If you found a refrigerator door in your yard on the morning of May 25, it may have been ours.

Most of the automobiles and trucks on our street did not vanish due to their large weight. They were merely picked up, bounced, and rolled for half a mile (Fig. 17).

Our two weenie dogs, Heidi and Maggie, had been left inside the house in their steel wire kennel. After considerable digging—during which time our little friends did not respond to our calling their names—we found them, bewildered and scared but unscathed. They were probably the only witnesses who went through the tornado above ground and emerged unhurt. I wish they could tell us what they saw and heard as the intense wind and water deluge and flying debris bore down on the house,

smashed the windows, ripped off the roof, and the walls crashed around them in a turbulent blizzard of sheetrock fragments, 2×4 s, and muddy insulation. Heidi and Maggie now carry a passionate dislike of thunderstorms.

Scratching around in dusty debris gives one time to philosophize about our relationship with material things. A few days after May 24 Rhonda sadly reflected that, despite our old house's flaws, she would miss it because now our children would not be able to show *their* children the house in which they were raised. A tornado makes history, but history lost must be counted among its casualties. At the same time, Rhonda always wanted to design her own house, and now she can do so. New beginnings can be good.

Standing Bear is right: Railing at the storm would be an exercise in futility. The out-of-equilibrium conditions that make storms possible also make life possible. That is the price of admission to Life—but it's well worth it.

It says something about the human spirit, and our perceptions of risk analysis, that new houses are being built on most of the sites of the ones that were swept away. Ours will be among them. Although our tornado was a harsh reality, there were some things of the spirit that it could not touch. Nor did it have the power to put us in bondage to fear. For that we have our many friends to thank.

Amid the dust and debris, we learned to feel the beauty. We felt it in the awesome power of nature. We felt it in the kindness of so many gracious people, the deepest beauty of all.

Appendix: In Appreciation of Complexity

Let us examine the differential equations that meteorologists try to solve when studying the time evolution of weather systems. The main ingredients are Newtonian fluid mechanics applied in a rotating reference frame, and thermodynamics.

When $\mathbf{F} = m\mathbf{a}$ is applied to an infinitesimal fluid parcel of mass $dm = \rho dV$, where ρ denotes the fluid's density and dV its volume, the result is the Navier-Stokes equation. Let us begin with

$$d\mathbf{F} = \rho dV (dv/dt). \quad (\text{A1})$$

We need to list the forces that act on the parcel, the terms that go on the $d\mathbf{F}$ side of Eq. (A1). Besides gravity $\mathbf{g}dm$, neighboring fluid elements exert contact forces on our parcel, normal and tangent to its surface. The force per unit area normal to the parcel's surface is pressure P , and frictional forces due to viscosity are tangent to the surfaces (recall Fig. 6). These contact forces are expressed through a "stress tensor" T_{ij} that carries the dimensions of force per area, or energy density. In terms of the stress tensor, the i th component of the contact force is written

$$(dF_i)_{\text{contact}} = T_{ij} n_j dA$$

where j is summed over all components, and n_j denotes the j th component of the outward-pointing vector \mathbf{n} that is normal to a parcel's patch of surface having area dA . Modeling the viscous forces as proportional to the shear (a "Newtonian fluid"), the stress tensor takes the form

$$T_{ij} = -P\delta_{ij} + \mu \left[\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right], \quad (\text{A2})$$

where μ denotes the coefficient of viscosity and δ_{ij} the ij th component of the unit matrix (the Kronecker delta symbol, equal to 1 when $i = j$ and 0 when $i \neq j$). When the gravity and surface forces are put into Eq. (A1), after using some vector identities and dividing out the dV one derives the Navier-Stokes equation,

$$-\nabla P + \rho \mathbf{g} + \nabla(\mu \nabla \cdot \mathbf{v}) + \mu \nabla^2 \mathbf{v} = \rho \frac{d\mathbf{v}}{dt}. \quad (\text{A3})$$

To make this even more interesting, we note that \mathbf{v} is a field, a function of position and time, so that by the chain rule for derivatives its time derivative may be partitioned to distinguish rates of change at a given location from rates of change due to transport, or "advection." For the i th component of velocity,

$$\frac{dv_i}{dt} = \frac{\partial v_i}{\partial t} + \frac{\partial v_j}{\partial x_j} \frac{dx_j}{dt}, \quad (\text{A4})$$

with j summed over all components. Equation (A4) is the i th component of the vector equation,

$$\frac{d\mathbf{v}}{dt} = \frac{\partial \mathbf{v}}{\partial t} + (\mathbf{v} \cdot \nabla) \mathbf{v}. \quad (\text{A5})$$

With the use of vector identities on the advective term of Eq. (A5), one can also write the acceleration in the form

$$\frac{d\mathbf{v}}{dt} = \frac{\partial \mathbf{v}}{\partial t} + [(\nabla \times \mathbf{v}) \times \mathbf{v}] + \nabla \left(\frac{v^2}{2} \right), \quad (\text{A6})$$

which goes into the right-hand side of Eq. (A3). The familiar Bernoulli's equation, which states the work-energy theorem (per unit volume) for a fluid,

$$P + \rho gz + \frac{1}{2} \rho v^2 = \text{const.}, \quad (\text{A7})$$

emerges as a special case (for zero viscosity and no curl) of the Navier-Stokes equation.

Because of viscosity, the near-surface flow will have a very different velocity profile from the flows far from the surface. Thanks to shear between layers of fluid, or between colliding currents of air, or induced by temperature gradients in initially static air, the velocity vector can change drastically in magnitude and direction as one follows an imaginary line upward from a given location on the Earth's surface.

Other constraints must also be considered. In weather systems a given mass of air is only shoved about, not created or destroyed. Therefore when a given volume loses mass, a current of



Photo by Dwight E. Neuenschwander



Photo by Dwight E. Neuenschwander

Fig. 17. This Malibu (top) belonged to our next-door neighbor Kevin. It landed about half a mile from his driveway. This Class A motor home (bottom) belonged to Darin across the street. It was carried a quarter of a mile.

mass has to flow through its surface. These considerations lead to the equation of continuity,

$$\nabla \cdot (\rho \mathbf{v}) + \frac{\partial \rho}{\partial t} = 0, \quad (\text{A8})$$

whose integrated form says that $Av = \text{const.}$ for flow through a tube of variable cross section A . In the absence of sources or sinks, a fluid speeds up when it goes through a constriction.

Newton's second law is intended to be applied in an unaccelerated frame, but we observe the weather from aboard a rotating Earth. Therefore we still have to transform Newton's second law into a rotating reference frame to take our planet's rotation into account. Suppose we apply $\mathbf{F} = m d\mathbf{v}/dt$ to the motion of a particle in an inertial frame, then transform the coordinates to those of a reference frame that rotates with angular velocity $\boldsymbol{\omega}$ relative to the inertial frame. Not only may a particle's location \mathbf{r} change in time with respect to the rotating frame (e.g., when an ant walks across a carousel floor), but \mathbf{r} also gets "carried around" by the rotation as seen from the inertial frame (the ant's location on the carousel gets carried around relative to the ground):

$$\left(\frac{d\mathbf{r}}{dt}\right)_{\text{inertial}} = \left(\frac{d\mathbf{r}}{dt}\right)_{\text{noninertial}} + (\boldsymbol{\omega} \times \mathbf{r}). \quad (\text{A9})$$

When we apply this result to $d\mathbf{v}/dt$ by starting in the inertial frame and transforming into the rotating frame coordinates, and assuming that $\boldsymbol{\omega}$ does not change in time (which would introduce additional terms), the acceleration picks up the terms $2\boldsymbol{\omega} \times \mathbf{v}$ and $\boldsymbol{\omega} \times (\boldsymbol{\omega} \times \mathbf{r})$, where the \mathbf{v} and \mathbf{r} now denote, respectively, the particle's velocity and location relative to the origin of the rotating frame. Transpose these terms over to the "F side" of Newton's second law, where they join the inertial frame's \mathbf{F} to become the Coriolis force $-2m(\boldsymbol{\omega} \times \mathbf{v})$ and the centrifugal force $-m[\boldsymbol{\omega} \times (\boldsymbol{\omega} \times \mathbf{r})]$.

At latitude λ the Earth's spin angular momentum vector has a vertical component $\omega \sin \lambda$. Thus the magnitude of the locally horizontal component of the Coriolis force on a particle of mass m is $2mv_h \omega \sin \lambda$, where v_h denotes the magnitude of the particle's horizontal velocity.

For Earth's rotation, $|\boldsymbol{\omega}| = 2\pi$ radians/day, so that $\omega^2 \sim 10^{-10} \text{ s}^{-2}$. Compared to gravity, the pressure gradient, and Coriolis force, the centrifugal force is typically negligible in large-scale weather patterns.

We also need thermodynamics with equations of state before we can solve mathematical models of weather systems. The simplest equation of state for the atmosphere assumes an ideal gas, so that

$$P = R\rho T, \quad (\text{A10})$$

where R denotes the gas constant modified to units of J/K kg instead of J/K mole, ρ the air density, T the absolute temperature, and P the pressure. Other useful thermodynamic results include those of adiabatic expansions, which for an ideal gas includes expressions such as $PV^\gamma = \text{const}$. We also need the latent heats for phase changes and relations that constrain them (e.g., the Clausius-Clapeyron equation, which describes dP/dT in phase changes).

We now have the equations that, when supplemented with boundary conditions, determine in principle the velocity field uniquely. However, due to nonlinearities in the coupled equations and due to turbulence whereby the concept of velocity as a continuous field breaks down, analytic models of meteorological systems can be solved only for the simplest of geometries, limiting their applicability for real-world situations. To make realistic quantitative progress, one must set up the equations and their initial conditions on a supercomputer, then step through them in small time slices to work out the consequences numerically. [10] Even then, due to the weather being a "chaotic" system (strong sensitivity to initial conditions, which means that a tiny change in them produces a drastically different outcome due to nonlinearities), one can tolerably predict weather patterns only a few days in advance.

To understand approximately the weather's main features that lead to atmospheric currents, supercells, and tornadoes,

we leave aside the full-blown sets of equations and take as our starting point a simplified version of the Navier-Stokes equation transformed to a rotating reference frame attached to the Earth's surface:

$$-\frac{1}{\rho} \nabla P + \mathbf{g} + \mathbf{f} - 2(\boldsymbol{\omega} \times \mathbf{v}) = \frac{d\mathbf{v}}{dt}. \quad (\text{A11})$$

The velocity \mathbf{v} denotes that of an air parcel relative to the Earth-bound coordinate frame. The terms on the left side are the most relevant forces per unit mass. The first of these we call the "pressure gradient force," which by itself would direct the velocity from regions of high pressure to those of low pressure. The gravitational field \mathbf{g} by itself would pull the air down to the surface. Into the third force density \mathbf{f} I have stuffed the effects of viscosity, including shear. The fourth term is the Coriolis force, which for horizontal motion deflects the velocity vector with a force of magnitude $2\omega v_h \sin \lambda$ at latitude λ , directed to the right of the original \mathbf{v} in the Northern Hemisphere and to the left of \mathbf{v} in the Southern Hemisphere. For example, if the horizontal forces are balanced and the viscosity negligible, the horizontal wind speed in geostrophic flow follows at once from Eq. (A11):

$$v_{\text{geostrophic}} = |\nabla_h P| / 2\rho v_h \omega \sin \lambda. \quad (\text{A12})$$

With these forces and their conservation laws, along with thermodynamics, a few general conclusions can be drawn about normal weather behavior as described in the text.

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Bibliography

- T.A. Blair and R.C. Fite, *Weather Elements: A Text in Elementary Meteorology* (Prentice-Hall, 1965).
- Chief Luther Standing Bear, *My People the Sioux* (Houghton Mifflin, 1929; Univ. of Nebraska Press, 1975; Bison Books 2006).
- R.G. Barry and R.J. Chorley, *Atmosphere, Weather, and Climate* (Holt, Rinehart, & Winston, 1970).
- G. England, *Those Terrible Twisters* (Globe Color Press, 1987).
- D.J. Acheson, *Elementary Fluid Mechanics* (Clarendon Press, 1990).
- H.B. Bluestein, *Tornado Alley: Monster Storms of the Great Plains* (Oxford Univ. Press, 1998).
- R. Gonski, *Internet Thunder*,
<http://www.erh.noaa.gov/rah/education/eduit.html>.
- J. Haby, <http://www.theweatherprediction.com/>.
- National Oceanic and Atmospheric Administration website,
http://www.noaanews.noaa.gov/2011_tornado_information.html.
- M. Svenvold, *Big Weather* (Holt & Co., 2005).

Notes

- [1] Other estimates say that for any given spot in the Oklahoma prairie, the mean time between consecutive direct hits will be, statistically, about a thousand years (see Svenvold, p. 3). Of course, during each tornado event all sites in central Oklahoma are equally likely to be hit.
- [2] E. Lorenz, "Deterministic Nonperiodic Flow," *J. Atmos. Sci.* **20**, 130–141 (1963). See J. Gleick, *Chaos: Making a New Science* (Penguin Books, 1987).
- [3] H.R. Byers and R.R. Braham, *The Thunderstorm*, U.S. Government Printing Office, Washington, DC (1949).
- [4] 1 bar = 1000 mb = 10^5 Pa = 14.5 psi; more precisely, reference sea-level atmospheric pressure is 1013.25 mb = 14.7 psi.
- [5] For air rushing over an automobile or airplane the boundary layer is very thin (~mm). The avoidance of boundary-layer separation, with its ensuing turbulence, is why an airplane tapers down on the back of its fuselage and wings.
- [6] During the phase change itself the thermodynamic process is briefly not adiabatic, just as the spark plug igniting fuel in a gasoline engine means that the combustion event itself is not adiabatic. In both cases, however, the expansion that immediately follows the combustion *is* adiabatic because the new shot of internal energy does not escape the system as heat. The combustion or latent heat release are seen as events, not processes.
- [7] <http://www.tornadochaser.net/cape.html>
- [8] For comparison, the energy density of TNT is about 7.5 MJ/kg according to Ref. 6.
- [9] Data from the NOAA website mentioned above. The website goes on to say, "The deadliest tornado on record in the United States was on March 18, 1925. The 'Tri-State Tornado' (MO, IL, IN) had a 291-mile path, was rated F5 based on a historical assessment, and caused 695 fatalities."
- [10] See Bleustein, Ch. 3, "Numerical Simulations Come of Age," for a qualitative description of numerical results applied to supercells and tornadoes.