Storm Systems

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by

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Introduction

The word "storm" is generally defined as a disturbance in the weather. In the context of hydrological science, the most obviously relevant type of weather disturbance is one that produces precipitation. The deposition of precipitation is the result of a number of processes, but the most basic requirement for precipitation is the ascent of air that contains water vapor. Ascending air cools by adiabatic expansion, and condensation begins when this cooling results in relative humidities at or near 100 percent (see HSA031). This condensation produces clouds that can develop precipitation following continued ascent of the air [see Lamb (2001) for a more complete description of how precipitation is formed]. Once it begins, the instantaneous rate of precipitation, R, is roughly proportional to the product Ewq, where E is the efficiency at which water vapor is converted to precipitation that reaches the surface, w is the ascent rate, and q is the water vapor mixing ratio (the mass of water vapor per unit mass of air) in the ascending air (Doswell et al. 1996). When the average precipitation rate, $\overline{R}$, is multiplied by the duration of the precipitation, D, the result is the total accumulated precipitation during the storm. Typically, the efficiency of precipitation production by storms is not particularly high, perhaps being on the order of 50 percent, occasionally much more, and often much less, and can vary from one storm to the next, as well as during the life cycle of a single storm. Precipitation efficiency depends strongly on the environment in which a storm occurs – dry environments increase the likelihood that precipitation will evaporate before it reaches the surface. Strong variation of the horizontal wind with height, called *vertical wind shear* is also thought to reduce precipitation efficiency.

The water vapor needed for precipitation is present in the air mostly as a result of the process of *evapotranspiration* – the combined effects of evaporation (see HSA048) from open liquid water (fozen forms of water do not release much water vapor) and transpiration of water vapor by vegetation (see HSA045). However, precipitation does not always fall at or near the place where water vapor is first introduced into the air. Rather, the moving air, or *wind*, often transports that added water vapor away from its source before it ultimately falls out somewhere else as precipitation. Therefore, even nonprecipitating weather systems are hydrologically pertinent, because they are associated with the atmospheric water vapor *transport* component of the hydrologic cycle. As a result of this transport, precipitation can fall in geographic locations that have little or no local evapotranspiration (e.g., the polar regions, or deserts), for lack of vegetation and/or open liquid water. The movement of air also alters the transpiration *rate*, so any discussion of hydrologically important storm systems must also consider those processes that control the wind.

Storm systems operate on a variety of spatial and temporal scales, and that is the basis for an orderly consideration of them herein. In order to understand storm systems, it is essential to know the basic phyics that govern atmospheric motion. Air within the lower atmosphere is a mixture of many gases, but is mostly nitrogen (about 78%), oxygen (about 20%), argon (about 1%), carbon dioxide (less than 1%), and water vapor (variable, 1% or less). Of these gases, only the water vapor content varies much within the lower atmosphere. The fact that water vapor is a constituent gas that, unlike the others, can change phase to liquid or solid forms within the range of temperatures and pressures found within the atmosphere is a critical factor in the evolution of weather systems. The magnitude of this contribution to the total energy of storm systems by water substance is due to the relatively high *latent heat* of water. When condensation of water vapor occurs, latent heat is released, often enhancing the processes leading to upward air motions that initiated condensation in the first place. Storms systems on the Earth are stronger than they would be if the planet was dry and without significant latent heat release from condensation (like the atmosphere of Mars). Clouds produced by condensation also influence strongly the local radiation balance and so alter the spatial temperature distribution. Incoming solar radiation is the ultimate energy source for all weather, and clouds are important factors in the local energy budget.

Atmospheric processes are governed by a complex set of equations that are the mathematical expression of physical conservation laws: the conservation of momentum (Newton's laws of motion), the conservation of energy, and the conservation of mass, including the total mass of water in any form. Also included is a thermodynamic equation of state – for most practical purposes, the atmosphere is well-approximated as an ideal gas. The complete set of equations (see, e.g., Holton 1992) describing atmospheric motions can be simplified in different ways, depending on the temporal and spatial scale of atmospheric processes under consideration. By reducing the complexity of the governing equations, it is possible to gain a qualitative understanding of those storm systems associated with a particular scale of motion. Although the real atmosphere makes no such simplifications, the dominance of certain processes at specific scales is simply a result of the integrated dynamics. The fact that the dynamical system describing the atmospheric is substantially nonlinear means that a quantitative treatment of the atmosphere is only possible via computer simulations that are necessarily only approximations to the mathematical equations. These mathematical equations are, in turn, only approximations of the real atmosphere. Nonlinear dynamics is why weather forecasting is so widely recognized as challenging and subject to considerable uncertainty (Lorenz 1993). Since our understanding of the atmosphere is incomplete and our measurement of atmospheric variables is neither perfectly accurate nor at infinite spatial and temporal resolution, forecasting storm systems is never going to be perfect. This uncertainty is tied to the nonlinearity of the dynamical system and so cannot be circumvented at any time in the foreseeable future.

The vertical structure of the atmosphere is broadly described in terms of layers. From the surface to a height of about 10 km, temperature decreases with height at a rate of roughly 6K km^{-1}. At a height of 10 km, the pressure has fallen to about 30 percent of the surface value. This layer from the surface to roughly 10 km is called the *troposphere*, the actual height of which varies, being generally lowest in the polar regions and deepest near the Equator. Above the troposphere is the *stratosphere*, within which the temperature generally increases with height up to around 40-50 km, where the pressure has fallen to less than one percent of the surface value. The boundary between the troposphere and the stratosphere is called the *tropopause*. Generally, most of what we call "weather" is confined within the troposphere and lower portions of the stratosphere. The depth of the troposphere compared to size of the Earth is comparable to the thickness of the skin on an apple. Thus, the complexities of storm systems are mostly confined to a very thin layer, and the fact that the atmosphere is so thin relative to the planetary scale is an important factor in the dynamics of large scale storms.

Large Scale Storm Systems

Seen from space, the cloud patterns (Fig. 1) show that cloudy regions in middle latitudes are broadly associated with relatively large disturbances that rotate cyclonically – that is, counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere. The main large-scale weather systems of middle latitudes are called *extratropical cyclones*, to distinguish them from tropical cyclones (see HSA030), which have very different dynamics. Outside of the tropics, the dominant contributions to the dynamics of the atmosphere are: gravity, planetary rotation (manifested by the so-called *Coriolis Force* that varies from a maximum at the poles to zero at the Equator), the sphericity of the Earth, the character of the topography (notably, the mountains and oceans), and the unequal distribution of temperature resulting from solar heating. Because storm systems of large scale within the troposphere are very flat, the airflow in such systems, therefore, is predominantly horizontal and the vertical motions are so weak (on the order of a few cm s^{-1}) as to defy accurate routine measurement.

Broadly speaking, the fact that the Earth is a sphere means that incoming solar radiation per unit area is always largest in the tropics, decreasing as one moves poleward. This radiational imbalance is the key factor in large-scale weather of mid-latitudes (see HSA026). The tilt of the Earth's axis, combined with the rotation around the Sun, produces the seasonal changes poleward from the tropics. Within the tropics, the effect of seasonality is much reduced compared to middle and polar latitudes. Hence, although the Earth is always warmer in the tropics than near the poles, the temperature difference between the poles and the tropics changes with the season, being at its maximum in late winter and its minimum in late summer. Cyclonic storms operate on the energy available as a result of the temperature contrast between the poles and the tropics. The extratropical cyclones are, therefore, most intense during the cool season and at their weakest in the summer. A cyclonic disturbance transports cold air from the polar regions toward the tropics, and warm air from the tropics toward the poles, thereby acting to reduce the horizontal temperature contrast.

Another important factor in the temperature distribution is the difference between oceans and land surfaces. Water has a much higher specific heat than that of land, so a given amount of incoming solar radiation changes the temperature of water much less than that of a comparable land surface. This damps the magnitude of the diurnal and seasonal temperature cycles over the oceans compared to that over land, and creates *land/sea breezes* on a daily time scale and *monsoons* on a seasonal time scale. Moreover, the ocean basins include large-scale currents (such as the Gulf Stream) of their own that modulate the structure of the overlying air and, in turn, are influenced by the airflow over the oceans. The oceans and the atmosphere together are a coupled dynamical system (see HSA027).

The existence of a north-south (or *meridional*) temperature gradient results in, among other things, a band of strong westerly winds at a height of about 10 km (near the tropopause) in midlatitudes called the *jet stream*. The meridional temperature contrast between the Equator and the poles is not evenly distributed but tends to be concentrated in a relatively narrow zone called the *Polar Front*. As a result of all the complicating factors already mentioned, the Polar Front itself varies in location and intensity from day to day and generally migrates poleward in summer and equatorward in winter. The overlying jet stream is in fact tied dynamically to the strength and location of the Polar Front.

On large scales, it turns out that extratropical cyclones are the size they are because disturbances of that size are most efficient at transporting heat from the equatorial regions poleward (and, equivalently, transporting cold air from the polar regions equatorward). The leading edges of cold air masses traveling equatorward are *cold fronts*, whereas the leading edges of warm air masses traveling poleward are *warm fronts*. Extratropical cyclones also provide some *vertical* transport of heat, as warm air involved in the storm typically ascends, whereas the cold air descends. The effect of the vertical heat transport is to carry the excess heat from solar radiation at the surface upward. The weak vertical motions on this scale are efficient at vertical heat transport only because of the large size of the air masses involved. On the average, it only takes about 3 to 6 extratropical cyclones per hemisphere to maintain the observed mean hemispheric thermal structure in the face of the continuing unequal solar heating (Palmén and Newton 1969). Without extratropical cyclones, the tropics would become much warmer than they now are, while the polar regions would cool still more, perhaps to the point where the habitable portion of the Earth would be confined to a narrow strip in middle latitudes.

Generally speaking, extratropical cyclones have a life cycle (Fig. 2) that includes a time of development, during which the increasing kinetic energy associated the winds of the storm is drawn from the potential energy available from the unequal distribution of temperature. The life cycle of an extratropical cyclone unfolds over several days. Given that these storms generally move from west to east in middle latitudes, owing to the generally westerly flow in which they are embedded, a new extratropical cyclone passes through a given region on the order of every few days. This results in a cycle whereby locations along the track of such storms experience warming as the extratropical cyclone approaches and cooling as the storm passes. The extratropical storm track is generally along the average location of the Polar Front, which can sometimes persist in the same area for many weeks, or it can shift as the overall airflow pattern changes.

When an extratropical cyclone develops, its horizontal and vertical wind speeds increase – the kinetic energy of the cyclone increases as it draws down the available potential energy. The airflow tends to organize itself in "conveyor belts" (see Browning 1985) where the ascending currents become cloudy and the descending currents tend to be cloud-free, resulting in the characteristic cloud patterns associated with extratropical cyclones (Fig. 1). The development of the storm also increases its overall rotation rate, drawing the clouds into spiral patterns.

Generally, the air to the east of an extratropical storm is rising, as well as moving poleward. If this poleward-moving warm air is also moist at low levels because at some point in the past it has passed over a moisture source (e.g., the warm waters of the Gulf of Mexico), then the likelihood of precipitation is relatively large. On the other hand, if the source of the warm air is from a dry region (like the Sahara Desert), then the likelihood of precipitation is relatively small. Conversely, the equatorward-moving cold air to the west of an extratropical cyclone is usually sinking, so the weather tends to be relatively fair. Depending on the local topography, various exceptions to these tendencies can be encountered.

It is noteworthy that the Northern Hemisphere has several large mountain ranges and other regions of relatively high terrain, and also has most of the land mass on the planet. On the other hand, the Southern Hemisphere has a much higher percentage of its area covered by oceans, with only one major mountain range, the Andes Mountains of South America. This asymmetry is also reflected in the complexity of the pattern of cyclones in their respective hemispheres: Northern Hemisphere weather patterns are far more complex, on average, than those in the Southern Hemisphere, where westerly flow (also called *zonal* flow) predominates in upper levels. The presence of complex topography also influences the distribution of weather within passing extratropical cyclones. Where air flows toward rising terrain, it is forced upward, which can cause precipitation if that ascending air is also moist (see HSA033). Thus, the western slopes of the Americas, with their coastal mountain ranges facing predominantly westerly winds from the Pacific Ocean usually have abundant rainfall. When air flows downslope, it tends to dry out – deserts or semiarid regions are generally found to the east of the North American mountain ranges, for instance.

Mesoscale Storm Systems

The distribution of precipitation within an extratropical cyclone (as illustrated schematically in Fig. 2) can vary substantially from one situation to the next, depending on the availability of water vapor in the air currents, and from time to time within the life cycle of the storm system. Embedded within an extratropical cyclone are smaller storm systems that occur on an intermediate scale, often referred to as *mesoscale* storm systems. The horizontal scale of such storm systems ranges from about 100 km to about 1000 km and the time scale ranges from a few hours to a few days. The specific conditions that favor the development of mesoscale storms within an extratropical cyclone are modulated by the details of structure and evolution for that particular large-scale system. Whereas large-scale meteorology is dominated by extratropical cyclones, mesoscale storm systems include a larger variety of processes than observed at large scales in mid-latidudes. Hence, it is difficult to make broad generalizations about mesoscale storm systems. Although there are tendencies for certain weather patterns associated with extratropical cyclones (Fig. 2), exceptions to those tendencies are common, because the distribution of weather within an extratropical cyclone depends on many factors (see HSA028). Extratropical cyclones are always present in midlatitudes, because there is always the temperature difference between the poles and the equator to drive them, but the smaller scale storm systems created within them are only present intermittently. Mesoscale storms themselves generally act to remove whatever large-scale conditions caused them in the first place. These mesoscale storm systems are complex in their variety, since on this scale, virtually no dynamical factor is always negligible. Mesoscale weather is therefore the most complicated and, therefore, the least understood – see Houze (1993) for a discussion of clouds and precipitation systems across a range of scales. Another factor complicating the understanding of mesoscale processes is the lack of quantitative observations of the needed resolution. Mesoscale storm systems can be broken down in two broad classes: (1) those tied to some topographic feature such as large lakes, coastal areas, orographic features, etc. and (2) those resulting from inherently mesoscale internal atmospheric mechanisms (Emanuel 1986).

a. Topographically-driven mesoscale storm systems

Some examples of mesoscale storm systems associated with topography are lake-effect snowstorms, upslope precipitation, mountain precipitation systems, sea/land breeze systems, and so on. A critical issue in such storms is their intimate connection to processes on larger scales. For example, if we consider lake-effect snow storms (Fig. 3), the occurrence and location of the snow depends very much on the wind direction of the large-scale flow relative to the lake in question. These mesoscale snow events arise as cold air flows across warm water, becoming moist and unstable at low levels as a result of sensible and latent heat flux from the water. As a large-scale system moves by, the direction of the prevailing low-level winds changes. Change that large-scale flow and the lake-effect snow will cease in one place but may commence in another. Hence, during the passage of an extratropical cyclone, the distribution of lake-effect precipitation will evolve and different areas could receive heavy snowfall on different days. Since each extratropical cyclone is different, the mesoscale details will vary, but large lakes are fixed topographical features, so there are *preferred* areas for lake effect snow. Exactly which areas will be affected and at what time depends on the detailed structure of the particular extratropical cyclone.

Upslope rain events are another example of a topographically-driven mesoscale storm. As moist air is forced upslope, it condenses and forms first clouds and then rain. As with lake effect snows, the direction of the flow at large scales interacts with the topography to produce important weather. Upslope rain involves thunderstorms when the air flowing upslope is moist and unstable. Since the situations in which upslope thunderstorms form can persist for many hours, the result can be prodigious rainfall rates, up to 200 mm hr^{-1}, for extended periods

The fact that topographically-driven mesoscale storm systems result from an interaction between topographic features and large-scale storms systems make them somewhat easier to predict. The greatest accuracy in weather forecasting is generally associated with the largest scale systems, so that when armed with a detailed knowledge of topography, it can be fairly straightforward to anticipate the general character of the topographically-forced mesoscale features that will occur during the passage of large-scale weather systems. Forecasting such systems is still neither easy nor perfectly accurate, because in forecasting, the small details are inevitably very important. The fact that topographic features are fixed or change only slowly means that the mesoscale storm systems tied to them can be somewhat more predictable than the other category of mesoscale storm systems – those that are driven by instabilities associated with internal atmospheric processes.

b. Free mesoscale storm systems

Although extratropical cyclones, the dominant large-scale storm systems, are the size they are as a result of a known scale selection mechanism that maximizes their efficiency at transporting heat meridionally, no such dominant scale selection dynamic process is known for mesoscale storm systems in general. Curiously, the so-called *fronts* that are characteristic of extratropical cyclones have a somewhat ambiguous scale: *along* such a front, they are clearly large-scale processes, extending for 1000 km or so, but perpendicular to a front, the front itself is mesoscale, with characteristic widths for the frontal zone being 10-100 km. The dynamics of fronts are reasonably well-known, and frontal zones can include many complicated mesoscale structures that are important for modulating the weather but are only poorly understood. Fronts are an example of a mesoscale process driven by the dynamics of large-scale weather systems.

Mesoscale cyclonic storms called *frontal waves* are often found in association with fronts. These might at times represent the early stages of a developing extratropical cyclone or they might remain within the mesoscale size range, being relatively transient features that nevertheless can influence the weather during their mesoscale life cycle. Theory says that such mesoscale perturbations within an extratropical cyclone tend to be confined to near the surface, shallow, and have shorter life cycles than the extratropical cyclones in which they occur (Gall 1976). Note that fronts themselves can be important in development of mesoscale regions of ascending air, leading to precipitation. Mesoscale storm systems have vertical motions that can be 10-100 times as strong as those associated with extratropical cyclones – those vertical motions, therefore are a few tens of cm s^{-1} to perhaps 1 m s^{-1}. Such relatively strong ascent is concentrated in mesoscale regions and contributes to much higher precipitation rates than would be found on the scale of the typical extratropical cyclone. In some regions, at certain times of the year, these mesoscale storm systems produce heavy rain and snow falls, such as within the East China Sea region in the wintertime, during the so-called Baiu Front season (Ninomiya et al. 1988).

Although individual thunderstorms are small enough to be considered "small-scale" weather systems and, therefore, are considered in the next section, under certain circumstances (described in Maddox 1983), many individual thunderstorms become organized into what are called *mesoscale convective systems* (MCSs) – see Fritsch and Forbes (2001) for more details. In many cases, the thunderstorms are linked together into lines of individual storms that interact strongly with each other to produce a mesoscale system, examples of which is shown in Fig. 4. Such systems can persist for many hours, and sometimes even for days. They can be associated with very heavy precipitation as well as severe convective weather (discussed in the next section). MCSs arise in a variety of ways, but some of the largest and most persistent examples, given the special name of *mesoscale convective complex* (MCC), appear to have a preference for certain regions of the world (Fig. 5), which suggests that their occurrence might be linked to topographic effects. Note that many MCCs are observed in the tropics, as well as in middle latitudes.

Mesoscale features are often embedded within extratropical cyclones at any time of the year, including the winter. In addition to lake-effect snows, there can be mesoscale regions of intense snowfall that travel along with the extratropical cyclone, producing swaths of heavy snow (and/or freezing precipitation) along their tracks. Such bands of winter precipitation are typically on the order of a few hundred km in width (Fig. 6), and so represent the track of traveling mesoscale winter precipitation maxima that are associated with processes within a much larger extratropical cyclone.

Within the wintertime polar air streams often can be found the so-called *polar lows* (Rasmussen and Turner, 2003). These are mesoscale storm systems that take on two rather different forms: when occurring over the ice-free ocean waters, some have been found to have many characteristics in common with tropical cyclones, including relatively weak horizontal temperature contrasts, warm cores, and occasionally even cloud-free “eyes”, despite their occurrence at high latitudes. Other types of polar lows are clearly small cousins to the extratropical cyclone, forming in association with strong horizontal temperature contrasts, sometimes associated with topographic features. Both can produce intense snowfalls in association with localized high winds.

Small-scale Storm Systems

As the scale of atmospheric phenomena decreases below that of the mesoscale, it again becomes possible to make simplifying approximations. For example, the curvature and rotation of the Earth can be neglected for many small-scale phenomena. Events that might plausibly be called storms on this scale are primarily *thunderstorms*. Thunderstorms arise when the input of latent and sensible heat at low levels cannot be moderated fast enough by processes on larger scales (Doswell 2001). Thus, the occurrence of thunderstorms and the weather they produce (strong winds, hail, tornadoes, extreme rainfall rates, and lightning strikes) is associated mostly with land surfaces (Fig. 7), as illustrated by the global distribution of lightning. Thunderstorms typically develop during the daytime within the warm season.

Exceptions to this can be found, owing to topographic details, as in the relatively high thunderstorm frequencies over the warm waters of the Gulf Stream, east of the United States. When the low levels are warmed and the air contains sufficient moisture, the atmosphere is said to become gravitationally unstable – under such circumstances, plumes of heated air rising from near the surface reach condensation, becoming *towering cumulus* clouds (Fig. 8a). These evolve rapidly into mature thunderstorm clouds, called *cumulonimbus* (Fig. 8b). After maturity, ordinary thunderstorms dissipate (Fig. 8c). The entire life cycle of such a prototypical *thunderstorm cell* is roughly 20-30 min (Byers and Braham 1949), and most thunderstorms include a sequence of such clouds forming, moving through their life cycles, and dissipating. Thus, they are termed *multicell* thunderstorms and they can be rather ordinary, or they can become severe thunderstorms. The criteria for storms being called severe are generally arbitrary; in the United States, if a thunderstorm produces winds of 50 knots (25 m s^{-1}) or stronger, hailstones of diameter 3/4 inch (2 cm) or larger, or a tornado, it is deemed a severe thunderstorm (Galway, 1989). Heavy rainfall is not considered officially severe in the United States.

The physical process driving thunderstorms is buoyancy and that buoyancy is strongly dependent on latent heat release. Thus, for thunderstorms to occur, the heat content at low levels must be high relative to that in middle and upper levels (instability), some process to cause air from low levels to rise to a height where it becomes buoyant is needed (lift), and there must be sufficient moisture in the ascending air to maintain the buoyancy. Without any one of these three ingredients (moisture, instability, and lift), thunderstorms cannot form. Thunderstorms act to reduce the instability by redistributing the heat (sensible and latent) from low levels upward. Once they have accomplished the needed heat redistribution, then thunderstorm activity ceases.

Most thunderstorms are ordinary, in the sense that they don't produce phenomena that meet the criteria to be called severe thunderstorms. Perhaps a few percent of all thunderstorms become severe, with most of them only barely meeting the criteria. Out of all the severe thunderstorms, a small fraction (on the order of a few percent) of them are *supercell* thunderstorms. Almost all supercells (around 95 percent) produce one form or another of severe weather. A supercell develops under conditions that favor thunderstorms in the presence of strong vertical wind shear. Thunderstorms that develop in strong vertical wind shear transform that vertical wind shear into rotation about a vertical axis, at times visually evident in a spiral structure to the storm (Fig. 9). In the northern hemisphere, this interaction results in counterclockwise rotation concentrated in the so-called *mesocyclone*.[1] It is the presence of a deep, persistent mesocyclone that distinguishes supercells from ordinary thunderstorms. Owing to the organizing dynamics of mesocyclones, supercells can become quasi-steady and persist for several hours, much longer than the typical ordinary thunderstorm cell. They also can be responsible for the most violent severe thunderstorm phenomena: families of strong-to-violent tornadoes, hailstones with diameters exceeding 2 inches (5 cm), and wind gusts exceeding 65 knots (32 m s^{-1}). This is the result of dynamical processes that produce the supercell and enhance the vertical motions in such storms beyond that expected from buoyancy effects alone.

Thunderstorms also can become organized into lines of interacting thunderstorm cells, sometimes called *squall lines*. These often are large enough to be considered mesocale in extent and so are properly termed MCSs. However, this tendency for linear organization of storms persists into scales arguably near or below any particular arbitrary threshold separating "mesoscale" from "small-scale". Any time that thunderstorms can become organized, the potential for severe weather increases. Although many heavy rain-producing thunderstorms are not severe by official criteria used in the United States (see above), they can be organized to produce dangerous flash floods under the right

[1] Note that this term is something of a misnomer, since it seems to suggest a mesoscale process, but the horizontal scale of a mesocyclone is on the order of a few km, too small to be properly considered mesoscale.

hydrological circumstances. Generally, thunderstorms that produce heavy rainfall are associated with the repeated passage of mature thunderstorm cells over the same general region; this is the so-called *training effect*, illustrated schematically in Fig. 10. Training of thunderstorm cells results in localized heavy rainfall amounts and is by far the most common evolution associated with flash-flood producing storms. Supercells often produce large instantaneous rainfall rates owing to their intense updrafts, but they typically do not remain in one place long enough to create high rainfall totals. Nevertheless, when the rainfall rate reaches 200 mm hr^{-1}, which some supercells have attained, they can produce dangerous urban flooding in as little as 15 min (Smith et al. 2001).

Nearly stationary rainstorms can also develop when air flows upslope, as noted above. Sometimes, such rainstorms occur with little or no lightning, while at other times, upslope rainstorms are associated with considerable lightning and thunder. Whether they are thunderstorms or not, they can produce small-scale regions of very heavy rainfall when the upslope slow persists for many hours.

On rare occasions, some snowstorms are accompanied by lightning and thunder. In such events, the instantaneous snowfall rates can be quite high (instantaneous rates might be as high as 25 cm hr^{-1}, but such rates are not typically sustained for long). Although intense winter storms such as lake effect snowstorms and upslope snowstorms are likely to be more properly considered mesoscale events, the peak values associated with such storms might be less than 100 km in spatial scale and so represent small-scale events embedded within a mesoscale storm system.

Related Articles

HSA001, HSA027, HSA028, HSA030, HSA031, HSA032, HSA033, HSA042, HSA045, HSA048, HSA053, HSA054, HSA110, HSA130, HSA179, HSA180, HSA206, HSA211

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Figure Captions

Fig. 1. Full-disk image of the earth from the geostationary satellite GOES-12 on 11 March 1997, showing two major extratropical cyclones in the northern hemisphere's Pacific Ocean: one approaching the west coast of North America, and another immediately behind it, in the central Pacific. Frontal cloud bands spiral inward toward their centers in a counterclockwise rotation. Others can also be seen rotating the opposite way in the southern hemisphere. A bright band of thunderstorm clusters marks the intertropical convergence zone between the northern and southern hemispheric circulations. (NOAA image)

Fig. 2. Schematic evolution of an extratropical cyclone (the center of low pressure is indicated by the red "L") in the Northern Hemisphere, showing the development of a wave on the polar front (a - upper left), the amplification of that wave (b - upper right), the mature phase of the cyclone, which is beginning to "occlude" (c - lower left), and the beginning of the dissipation of the cyclone (d - lower right), as the occlusion process proceeds. Cold fronts are shown in blue, warm fronts in red, and occluded fronts in purple; dark green shading indicates regions of precipitation and light green shading indicates clouds; the black lines with arrows indicate the surface pressure with winds roughly parallel to the pressure contours. (Schematic drawings provided courtesy of NOAA)

Fig. 3. Visible satellite image (left) and radar-observed precipitation (right) associated with a lake effect snow storm, showing flow from the west-northwest across Lake Ontario, with multiple cloud bands producing snow over the eastern part of the

lake and on into the state of New York. (NOAA images provided by College of Dupage).

Fig. 4. A mesoscale convective system (MCS) along the southern coast of the United States as seen in an enhanced image from a geostationary satellite (left – NCAR image, used by permission) and a radar image of a similar MCS-associated line of thunderstorms (right – NOAA image). Note the size of the MCS affecting the states along the southern coast, compared to the extratropical cyclone system which fills the image, centered west of the Great Lakes. Radar reveals the locations of the strong convective cells (in white) that are powering the MCS, whereas satellite images show the high, cold cloud tops near the tropopause that cover the whole convective system.

Fig. 5. The global distribution of mesoscale convective complexes (MCCs), based on satellite imagery. Small squares indicate location of MCC at time of maximum extent; adapted from Laing and Fritsch (1997).

Fig. 6. Satellite image from 05 December 1999, 1815 UTC, showing a snow band of mesoscale width (roughly 100-150 km), from southeast New Mexico, across the Texas Panhandle, Oklahoma, and Kansas. The Canadian River valley can be seen as the dark line within the snow band in the Texas Panhandle, as can several large man-made lakes (the dark spots) within the snow band in Oklahoma and Kansas. (NOAA Image provided by the Storm Prediction Center)

Fig. 7. Global lightning flash distribution (from Christian et al., 2002).

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Fig. 9. Tornadic supercell thunderstorm on 3 June 1999, showing structures characteristic of a rotating storm, in which air in the storm is moving left to right in the foreground, into the photograph on the right-hand edge of the clouds, and right to left in the background, spiraling inward to the tornado, which is partially obcured by precipitation. (Photograph © C. Doswell, used by permission.)

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Figures

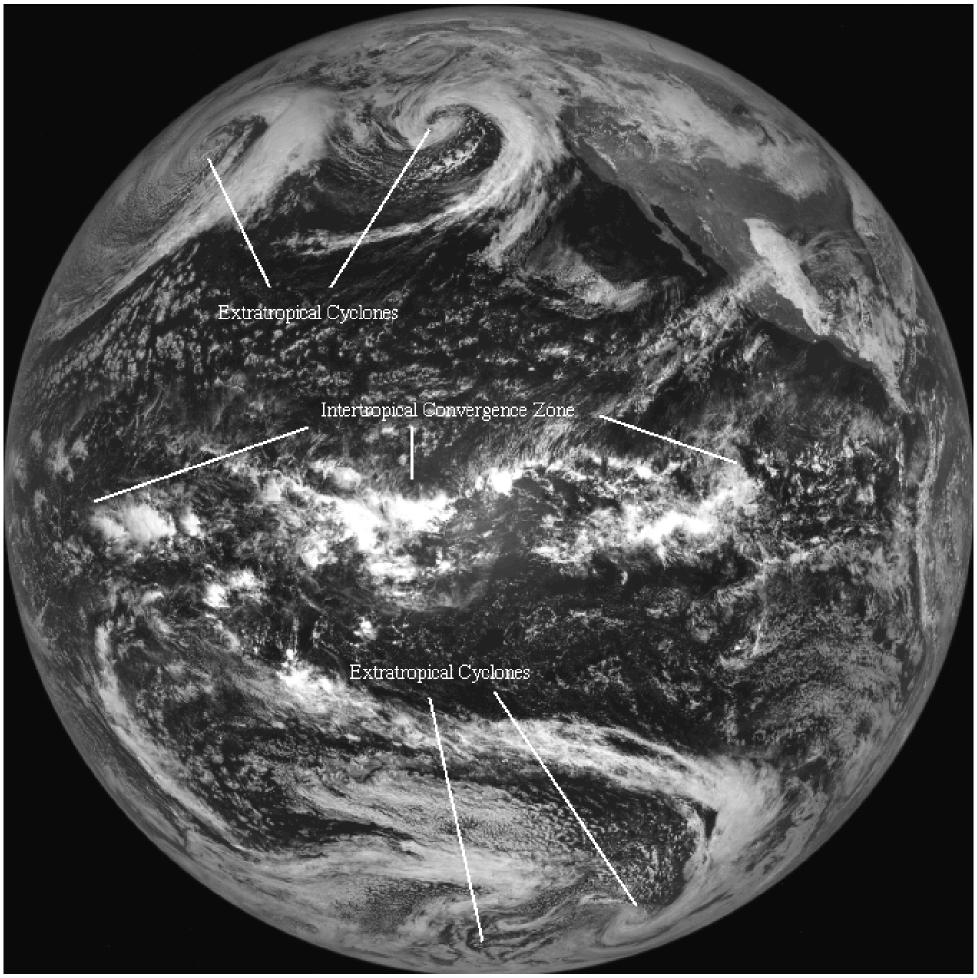
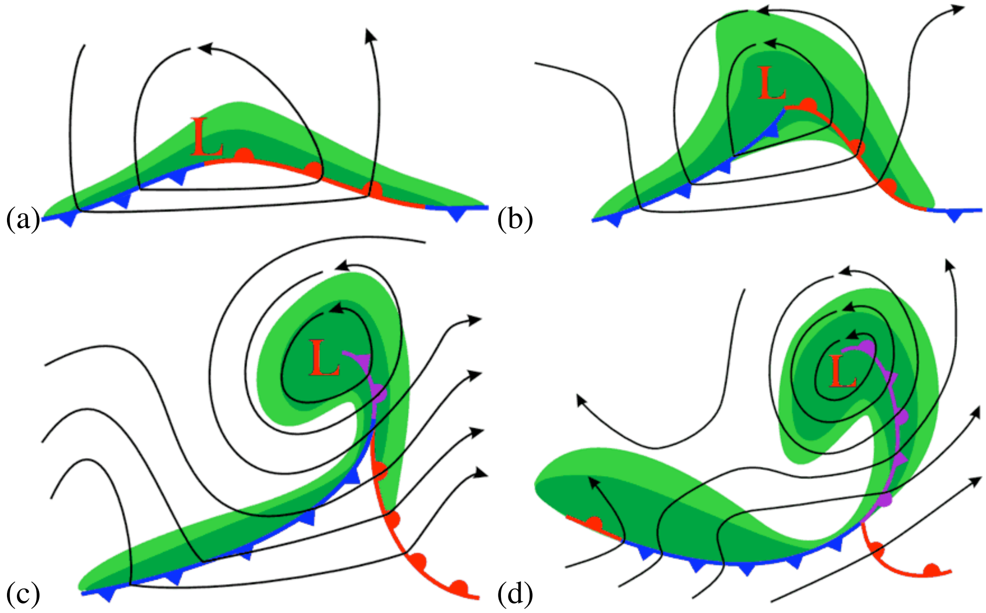


Fig. 1. Full-disk image of the earth from the geostationary satellite GOES-12 on 11 March 1997, showing two major extratropical cyclones in the northern hemisphere's Pacific Ocean: one approaching the west coast of North America, and another immediately behind it, in the central Pacific. Frontal cloud bands spiral inward toward their centers in a counterclockwise rotation. Others can also be seen rotating the opposite way in the southern hemisphere. A bright band of thunderstorm clusters marks the intertropical convergence zone between the northern and southern hemispheric circulations. (NOAA image)



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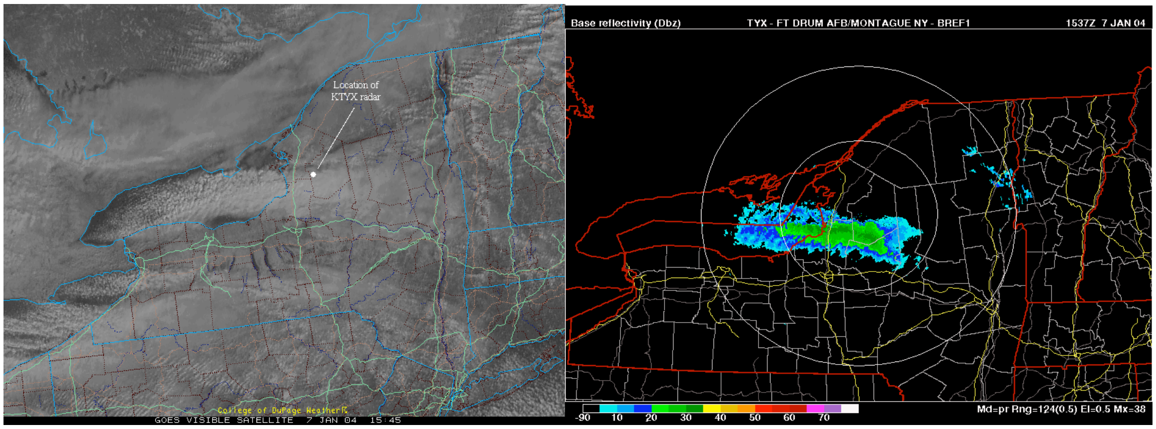

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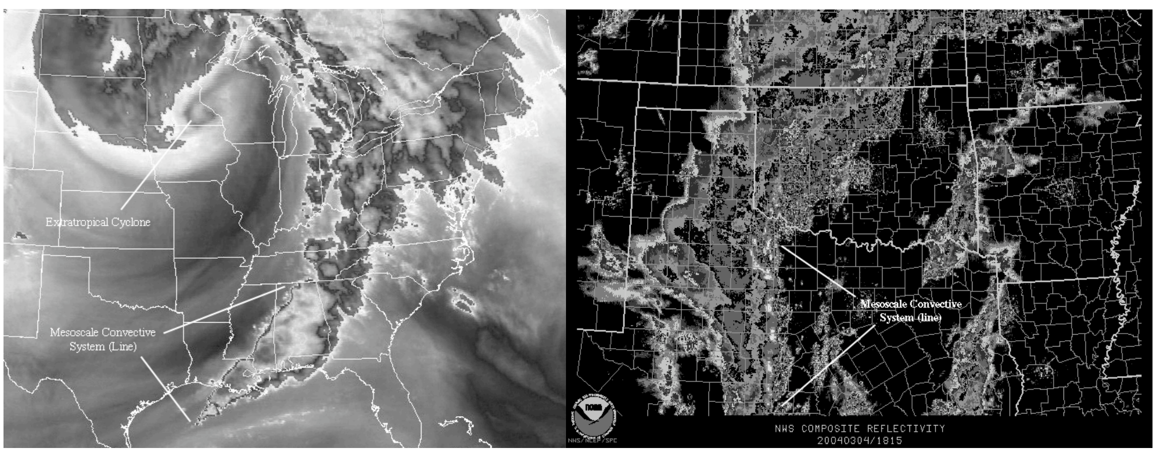

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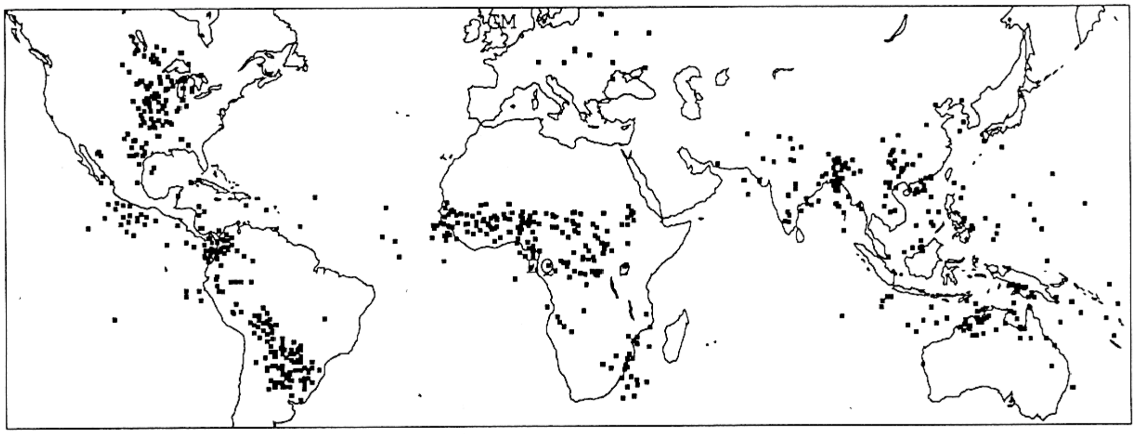

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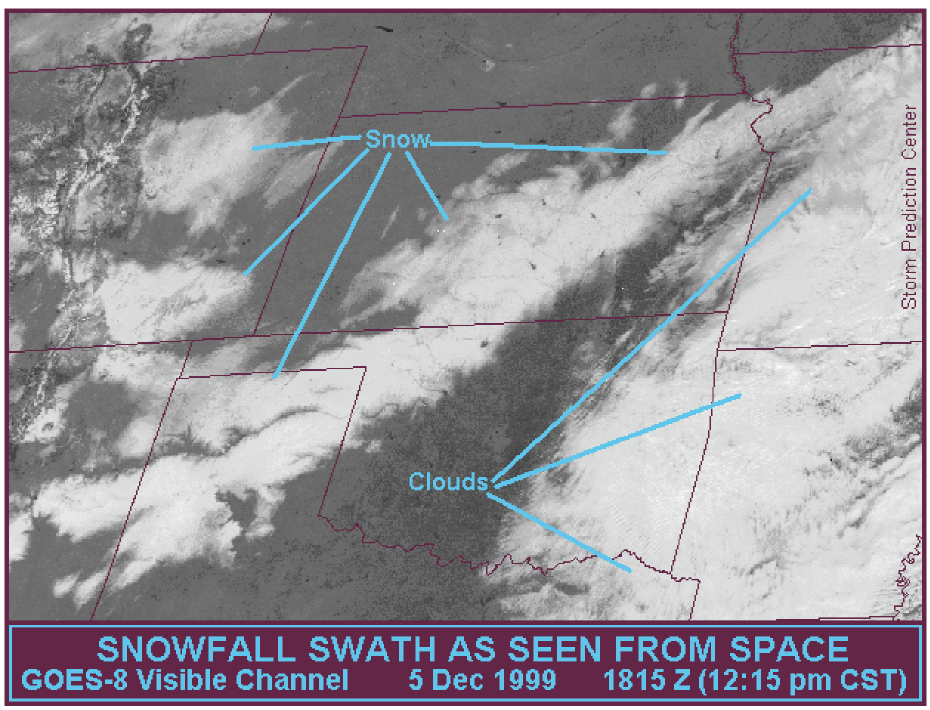

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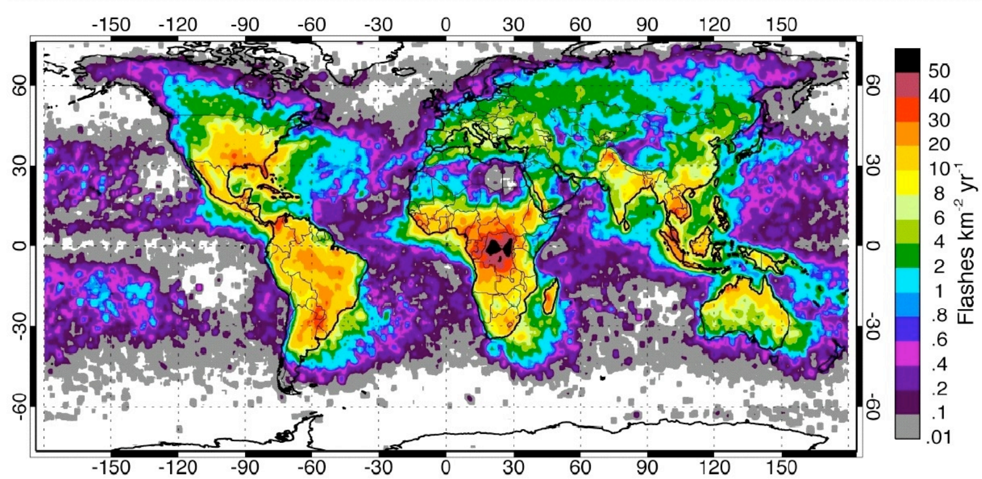

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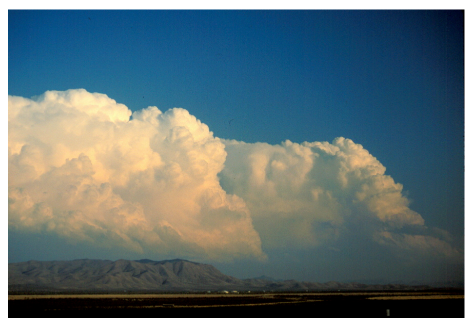
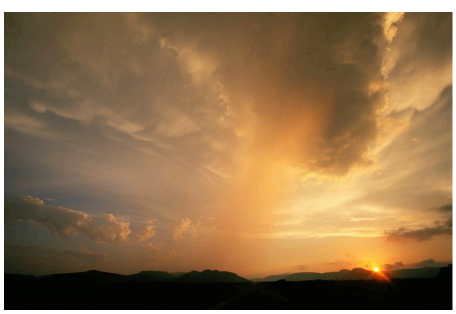
(a)  (b) 

(c) 

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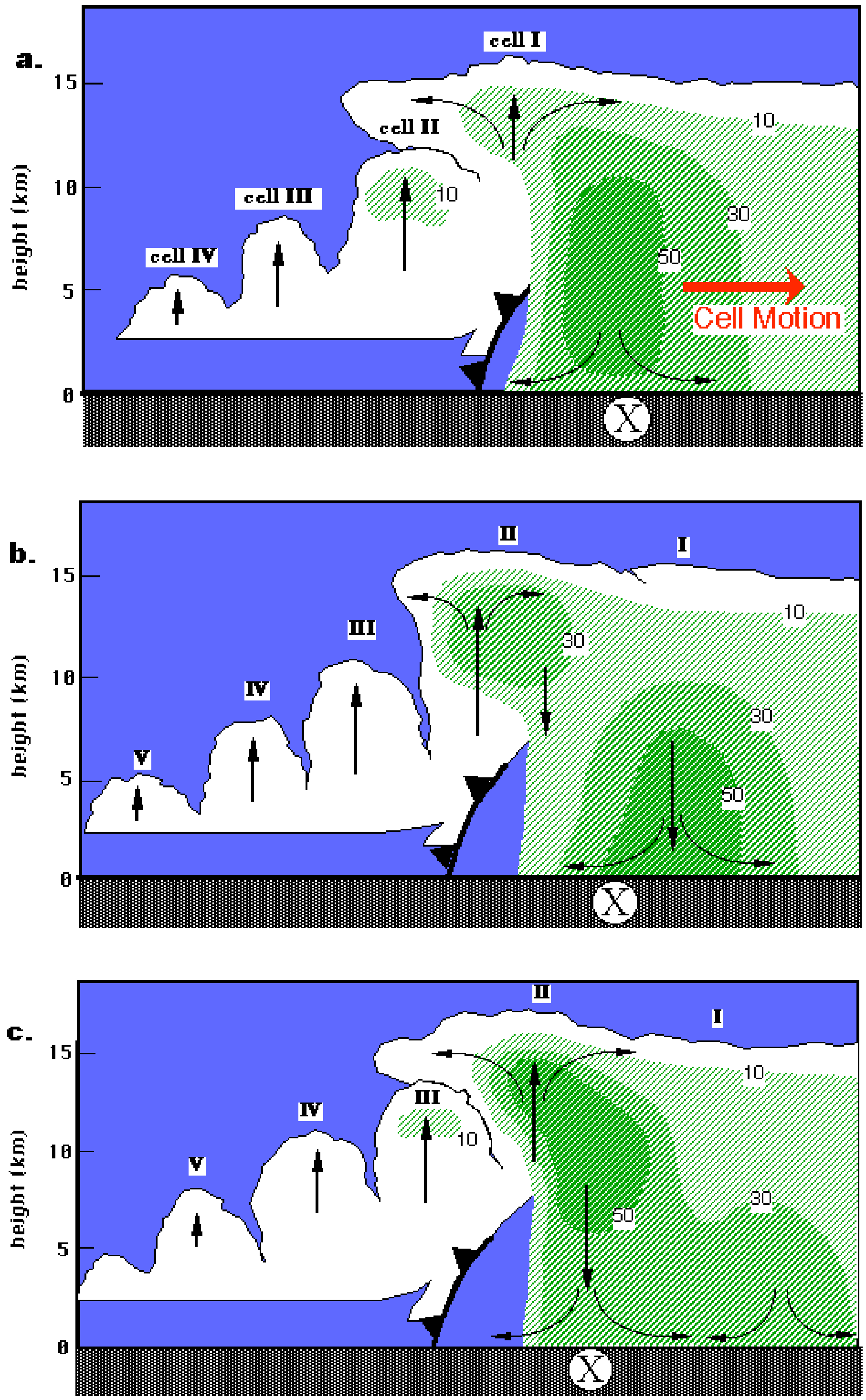

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