The Indian Monsoon

2. How do We get Rain?

Sulochana Gadgil

The most important facet of weather and climate in a tropical region such as ours, is rainfall. I have considered the observed space-time variation of the rainfall over the Indian region, in the first article[1] in this series. The ultimate aim of monsoon meteorology is to gain sufficient insight into the physics of this variation for predicting the important facets, with the help of atmospheric models based on the laws of dynamics and thermodynamics. Here I consider what we know about the rain-giving systems.

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[1] Part 1. Variations in Space and Time, *Resonance*, Vol.11, No.8, 2006.

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Clouds, tropical rainfall, monsoon rainfall, cumulonimbus.

Introduction

We know that we get rain from clouds. A cloud is defined as a visible aggregate of minute particles of water or ice, or both, in the free air. Most of the rain over the Indian region comes from the so called convective (i.e., Cumulus and Cumulonimbus) clouds. Cumulus is a Latin word for a heap or a pile, and cumulus clouds are generally dense with sharp outlines, developing vertically in the form of rising mounds, domes or towers of which the bulging part often resembles a cauliflower. The sunlit parts of these clouds are brilliantly white, their base is relatively dark and horizontal. Cumulus clouds are normally less than a kilometer in horizontal and vertical extent in their early stage of development. Most of them do not grow any larger, particularly when isolated. These are the fair-weather cumulus we see frequently from airplanes. A large cumulus cloud (Cumulus congestus *Figure* 1a) consists of a heap of rapidly fluctuating bulbous towers which give it its cauliflower like appearance. Cumulonimbus (nimbus means precipitating) which is an advanced stage of the development of cumulus cloud, is a heavy and dense cloud with considerable vertical extent, in the form of

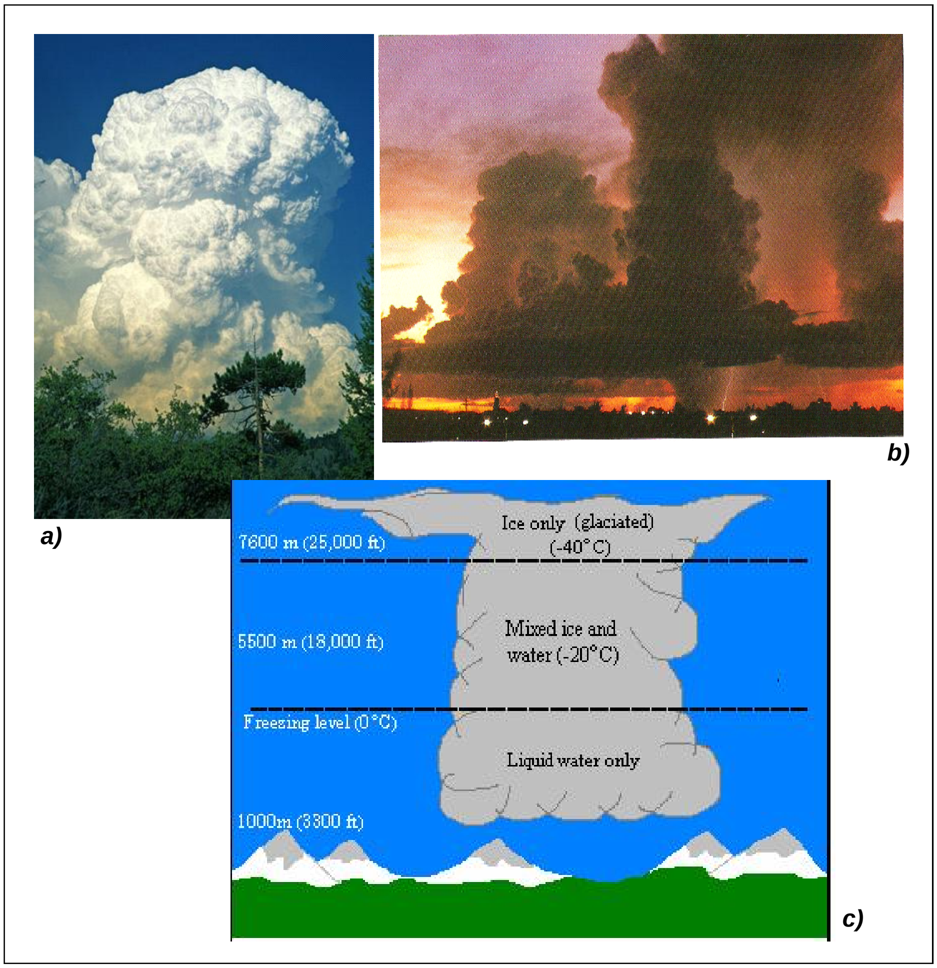


Figure 1. (a) Cumulus congestus. (b) Cumulonimbus. Notice the characteristic anvil on top. (c) Distribution of ice and water in a cumulonimbus cloud.

huge towers (*Figure* 1b). The tops are very high (well over 10km) and generally contain ice (*Figure* 1c). These rain-giving clouds are typically a few kilometers in horizontal extent. In the summer months of April and May we often get rain accompanied by thunder and sometimes also hail, from isolated clouds of this type.

Generally, the satellite picture of the tropical region on any day shows that the convective clouds are organized into systems of larger spatial extent viz. synoptic scale systems with horizontal scales of hundreds of kms (the lows, depressions and cyclones) and planetary scale cloud-bands extending over thousands of

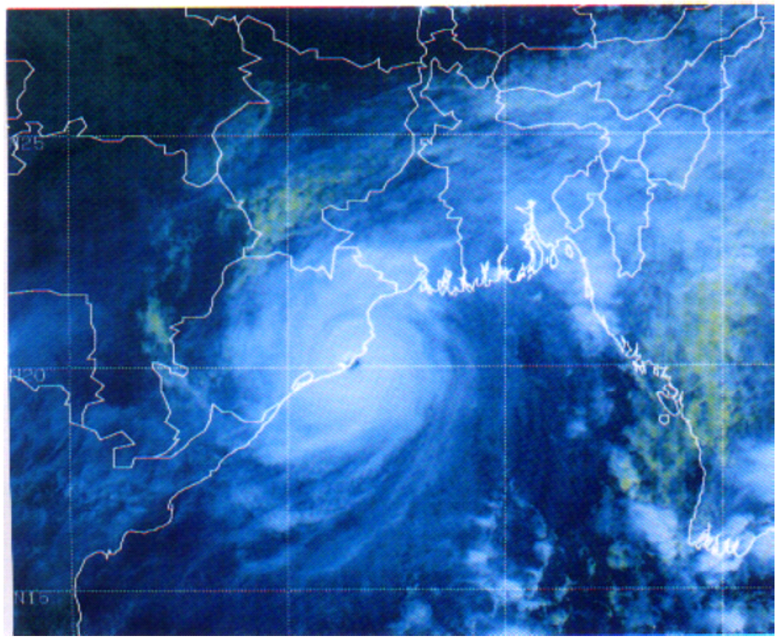
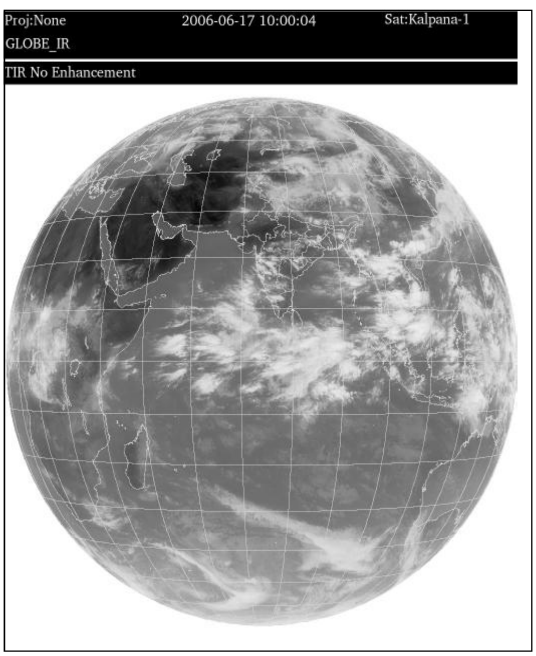


Figure 2. INSAT 1D image showing the landfall of super cyclonic storm of 29 October 1999 near Paradip on Orissa Coast.

kms (*Figures* 2 and 3). Most of the rainfall in our monsoon season, comes from such synoptic or planetary scale systems. We consider the processes that lead to (i) cumulus clouds in a tropical atmosphere and (ii) organization of hundreds of these clouds into synoptic and planetary scale systems in the following sections.

Figure 3.



Convection in the Tropical Atmosphere: Cumulus Clouds

The basic energy source for the atmosphere is radiation from the Sun. This incoming radiation is primarily in short waves, with a peak in the wavelength around 0.5 micron (which is one millionth of a meter) as the Sun is very hot (about 5800K). The composition of the atmosphere is such that the solar radiation passes almost unhampered to the surface of the earth where it is absorbed. So, the atmosphere is heated from below and hence the temperature of air decreases with height in the troposphere

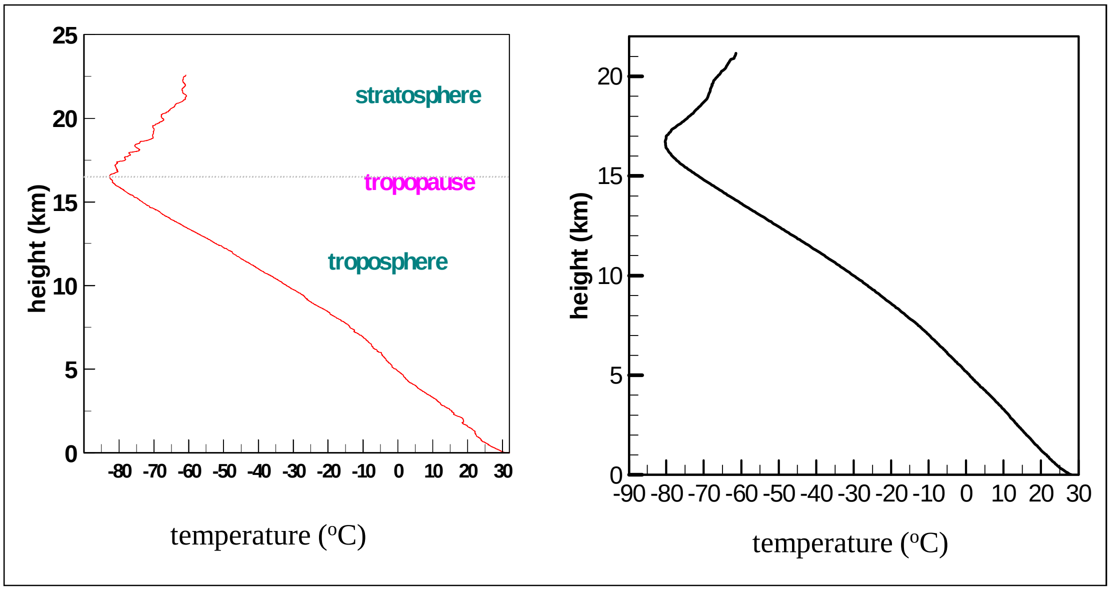
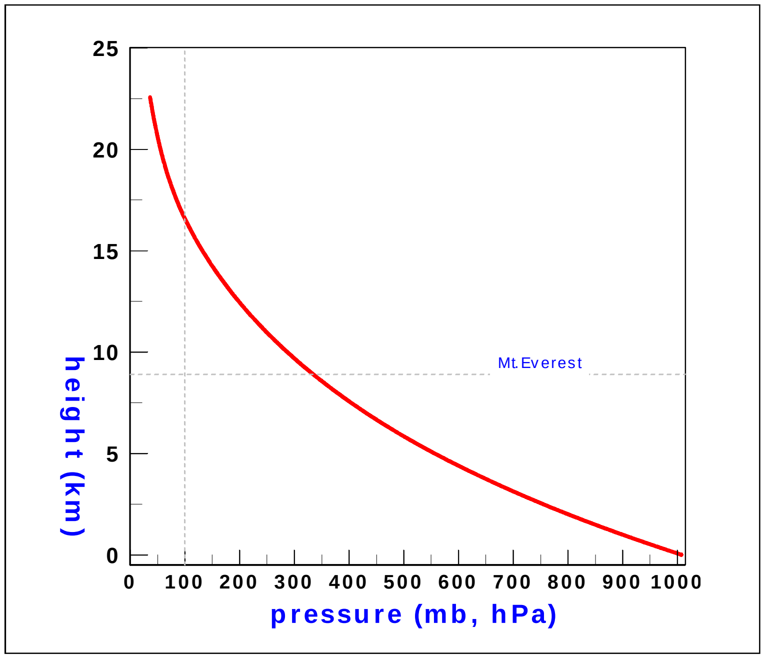


***Figure 4a.** Vertical variation of temperature with height over the Arabian Sea on May 28, 2002 (left) and over the Bay of Bengal during July August 1999 (right).*

which extends from the surface to a height of about 15-16 km in the tropics (*Figure* 4a). We are primarily concerned with the troposphere here, because it contains nearly all of the water and hence the clouds in the atmosphere. The earth–atmosphere system, in turn emits radiation but at longer wavelengths (peak around 10 microns), because it is cooler (257K). The associated net radiative cooling of the atmosphere is about 1.5 °C per day. The pressure at any point is simply the weight of the air column above the point, since the atmosphere is almost in hydrostatic balance [1]. Hence the pressure decreases with height (*Figure* 4b). The amount of water vapour in the air also decreases with height as it is supplied to the atmosphere through evaporation from the sea-surface. The ratio of the mass of water vapour in air in a given volume to the total mass of the moist air in that volume is called the specific humidity of the air. The amount of water vapour air can hold depends on the temperature of the air. Warmer air can hold more than colder air. The air is said to be saturated when it contains the maximum possible amount of water vapour for that temperature. The ratio of actual amount of water vapour in air to the maximum amount at that temperature (expressed as a percentage) is the relative humidity. The air near

Figure 4b. Variation of pressure with height



the surface of the tropical oceans is moist but rarely saturated, with the relative humidity generally around 80%. Since clouds contain liquid water, genesis of clouds involves a phase change of the water vapour present in the atmosphere. Such a change would occur when the air becomes saturated (i.e., relative humidity of 100%). The relative humidity of air increases if air cools. Generally such a cooling occurs in association with ascent of air. To understand how ascent leads to cooling, consider what happens to a parcel of moist air which is made to rise adiabatically (i.e., without giving it any additional energy) from a level A to a higher level, B, in the atmosphere. It will expand, because the pressure is lower at the level B and it will cool because this expansion is adiabatic i.e., at the expense of its own internal energy. As it cools adiabatically, the saturation vapour pressure decreases and at a certain level (depending on the humidity of the parcel at the start), it becomes saturated, i.e., relative humidity becomes 100%. Beyond this level, if the parcel is pushed up adiabatically, water vapour is converted to liquid water implying genesis of clouds. When water vapour is converted to liquid water, latent heat of condensation is released.

Thus for cloud formation, it is necessary for surface air to be lifted up to the level at which condensation begins (the *lifting condensation level*). Once the lifting condensation level is reached, it is possible for the parcel of air to keep going upward because of an instability of the tropical atmosphere called the *conditional instability*. A state is said to be unstable if when slightly perturbed, the small perturbation grows with time. Otherwise, the state is said to be stable (as discussed by Arakeri and Dharuman [1]). The tropical atmosphere is unstable with respect to vertical displacement of moist air but stable with respect to vertical displacement of dry air. This conditional instability can be understood as follows.

For cloud formation, it is necessary for surface air to be lifted up to the level at which condensation begins (the lifting condensation level.

A combination of the radiative, sensible[2] and latent heat transfer maintain the tropical atmosphere in a state of thermal equilibrium such that the temperature decreases with height at a rate of about 5.5° C per km in the lower part. It can be shown that the cooling associated with adiabatic ascent is at the rate of 10° C per km, i.e., at a rate faster than the environment. Hence it will be cooler and denser than its environment when it ascends adiabatically. Consequently, it will return to the lower level as it is acted on by a restoring buoyancy force. However, if we consider a parcel of moist air rising beyond the lifting condensation level, latent heat of condensation is released. This prevents it from cooling as rapidly as a dry parcel and in fact it cools at the rate of about 4.5 ° C per km which is lower than the environmental lapse rate of 5.5 ° C per km. Hence it finds itself warmer than the environment and is accelerated upward because of the buoyancy force. Thus the tropical atmosphere is unstable with respect to upward displacement of moist air parcels, beyond the level of saturation. Cumulus clouds are a manifestation of the gravitational instability and convective overturning in an atmosphere whose lowermost layer is moist.

[2] Sensible heat is that component of heat which can be sensed by humans i.e. that portion of the total heat associated with a temperature change. This is in contradistinction with latent heat which is hidden until phase changes occur such as when water vapour condenses to form liquid water.

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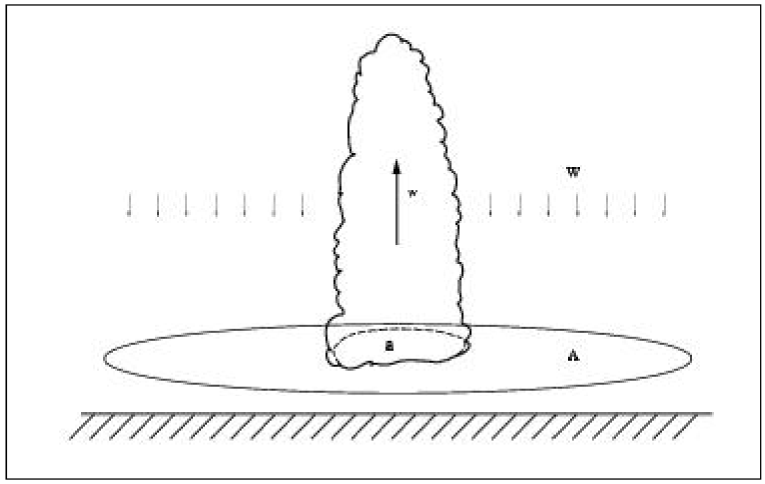
In order to get an idea of the resulting spatial scales of the circulation/convection patterns, we need to consider how the growth rate of the perturbations varies for the different modes/ spatial scales of perturbation. We expect that the mode/scale we

Figure 5.



will observe as a result of the instability process is the one that is the most efficient in tapping the conditional instability and hence has the fastest rate of growth. If the cumulus cloud is indeed a manifestation of the conditional instability of the tropical atmosphere, then convection with ascent of air over regions of the cumulus-scale (of a few km) has to be favoured over the larger scales in the presence of conditional instability. Bjerknes [2] first suggested how the cumulus scale would be selected for a conditionally unstable tropical atmosphere. Consider a cloud extending over a horizontal area '*a*'. We assume that dry air surrounding the cloud descends through the atmosphere over the surrounding region to maintain the mass balance. Conservation of mass requires that the ascent of air over '*a*' has to be balanced by descent over the surrounding region, which has an area '*A*' (*Figure* 5). Hence w, the velocity of ascent in the cloud is related to 'W', the velocity of descent in the surrounding region by

$$a \star w = A \star W.$$

Air rising in the cloud gains heat from the release of latent heat of condensation and loses heat because of adiabatic expansion, the rates of heating being dependent on the velocity of ascent, w. Since the tropical atmosphere is stable with respect to vertical displacements of dry air, work has to be done against gravity to force the descent of the dry air in the surrounding region. Thus, the buoyancy forces in the cloud have to do work to return the air pumped up by the cloud, to the surface. Consider a set of clouds of varying horizontal extent, with the same quantity of air involved in the vertical circulation for each member of the set. The smallest cloud (with smallest value of the ratio a/A) is associated with the maximum ascent velocity, w, of air in the cloud and the minimum velocity of descent, W, in the surrounding region. It can be shown that the ratio of the work done in the downward flow in the cloud-free region to the energy released by the buoyancy forces in the cloud is minimum for the cloud with maximum velocity of ascent and minimum area [3]. Thus the smallest horizontal scale of ascent is most efficient in tapping the instability and will be selected for.

Recent studies (Bhat [4] and references therein) suggest that over that large fraction of the tropical atmosphere which is cloudfree, descent occurs because of the radiative cooling of air which balances the heating associated with adiabatic compression. Thus, the descent of air around the clouds does not have to be driven by the buoyancy forces in the clouds working against gravity in a stable atmosphere as proposed by Bjerknes [2]. Rather, the downward flux of the air subsiding over a large part of the tropics due to radiative cooling is compensated by the pumping up of the air from the surface to the top of the troposphere by the deep clouds. In any event, the argument that the smallest scale will be selected holds in this case as well.

However, the selection for the cloud with the smallest area, holds only for an idealized fluid in which viscosity is ignored, so that the parcel of air rises without entrainment of surrounding air. In a real atmosphere the cloud will entrain air from the surrounding environment. Since this air is entrained at the boundaries of the cloud, the rate of entrainment (as a percentage of mass increase) will depend on the ratio of the boundary to the volume. For a cylindrical cloud this ratio per unit depth will be 2/radius.Thus the thinnest clouds have the highest ratio of surface to volume and will therefore have the most severe constraints of growth when effects of entrainment of the surrounding air due to viscosity are incorporated. However, it is known that these effects will not be important until the horizontal extent is of the same order as the depth of the troposphere. Thus, the selected scale of the cloud in a real atmosphere is not vanishingly thin, but of the order of 10 km. The typical horizontal extent of a cumulonimbus cloud observed in the tropical atmosphere is about 5 km and the typical life-span is about an hour.

We have seen that ascent of the moist air at the surface, at least up to the lifting condensation level, is a necessary condition for genesis of clouds. We get heavy showers on some afternoons in April–May, when the surface air is forced to ascend because of the intense heating of the land. The resulting thunderstorm, associated with an isolated cumulonimbus cloud, is often accompanied by hail since the cloud is deep with the top above the freezing level.

While the rainfall we get comes from convective clouds generated by ascent of moist surface air, we know that all such clouds do not produce rain. Cloud microphysics plays an important role in determining whether the cloud will rain. For rainfall to occur the droplet size has to be larger than about 30–50 microns. The drop size in turn depends on the size of condensation nuclei (i.e., salt particles over the oceans) and how deep the cloud is (it is large when top of clouds is above the freezing level). I will not consider cloud microphysics here, although it is obviously of great interest.

Convection Over Synoptic and Planetary Scales

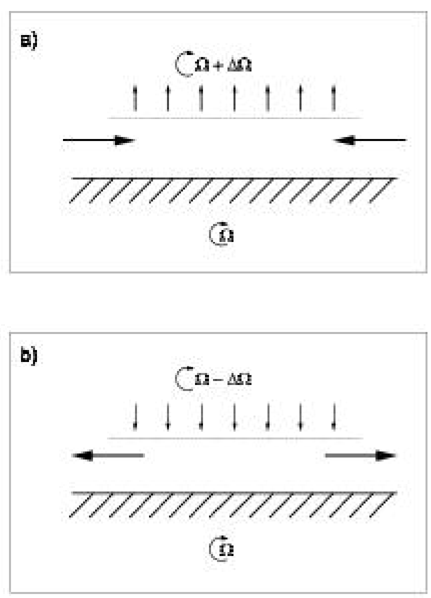
The ascent of moist surface air leading to genesis of cumulus clouds can also be a consequence of wind being forced to go over orographic features such as mountains. But in synoptic and planetary scale cloud systems (i.e., systems of spatial scales of hundreds or thousands of km respectively, e.g., *Figures* 2, 3), which are large enough for the rotation of the earth to become important in their dynamics, the ascent of surface air occurs

Box 1. Vorticity

Perhaps, the most important characteristic of fluid flow, is its vorticity. Vorticity is defined as the 'curl of the velocity' and is related to the rotation of the flow. Suppose that an infinitesimally small sphere of the fluid is instantaneously solidified (made rigid) without any change in momentum , angular momentum or mass distribution. Then the sphere will start rotating with an angular velocity which is half the vorticity of the flow. A simple way to see if a flow of a stream, for example, has vorticity is to let a cross (made with a pair of wooden sticks fixed at right angles) float along and see if it rotates. In the flow with parallel streamlines, the cross will rotate if there is shear, i.e. the velocity varies from one streamline to the next. Similarly, the cross will also rotate in a flow along a closed loop. In the northern hemisphere, the rotation of the earth (relative to which we observe the winds) is counterclockwise or what meteorologists call 'cyclonic'.

because of a very special characteristic of a boundary layer in a rotating system. In a rotating system when the flow in the interior of the fluid (where viscous effects are negligible) above the surface boundary layer has cyclonic vorticity (*Box* 1), i.e., rotation in the same sense as the rotation of the earth, there is convergence of air in the boundary layer giving rise to ascent above the boundary layer (*Figure* 6a). On the other hand when the vorticity in the interior is anticyclonic, the earth rotates faster than the air above the boundary layer, air diverges in the boundary layer leading to descent of air from the interior to the boundary layer (*Figure* 6b). In the tropical atmosphere, the boundary layer laden with water vapour, is typically about 1 km in depth. The synoptic and planetary scale convective systems (such as those seen in *Figures* 2, 3) are associated with cyclonic vorticity above the boundary layer.

Figure 6.



We need to note one more characteristic of the flow in a rotating system to relate the pressure patterns associated with these systems seen on weather charts, with the wind and the vorticity. The flow in a rotating system also differs from flow in a non-rotating system in another important feature, viz. the relationship between pressure gradients and flow velocities. We know that the flow in a water pipe is down the pressure gradient. However, in a rotating system when friction is unimportant, it is the Coriolis force which balances the pressure gradient. This implies that the flow is along contours of equal pressure (isobars), rather than at right angle to the isobars (i.e., down the pressure gradient). The Coriolis force acts in a direction perpendicular to the velocity vector and is to the right of the velocity vector in the northern hemisphere. Hence the flow around a low pressure area is counterclockwise in the northern hemisphere, i.e., cyclonic. The synoptic scale convective systems such as a depression or a hurricane, are characterized by closed vortices seen as closed isobars on the weather chart (e.g., *Figure* 7 for a depression on August 16, 2006). On the other hand, the planetary scale convective systems (such as the one associated with the cloudband in *Figure* 3) are generally associated with cyclonic vorticity arising from wind shear. While the flow in the interior above the boundary layer is along isobars,

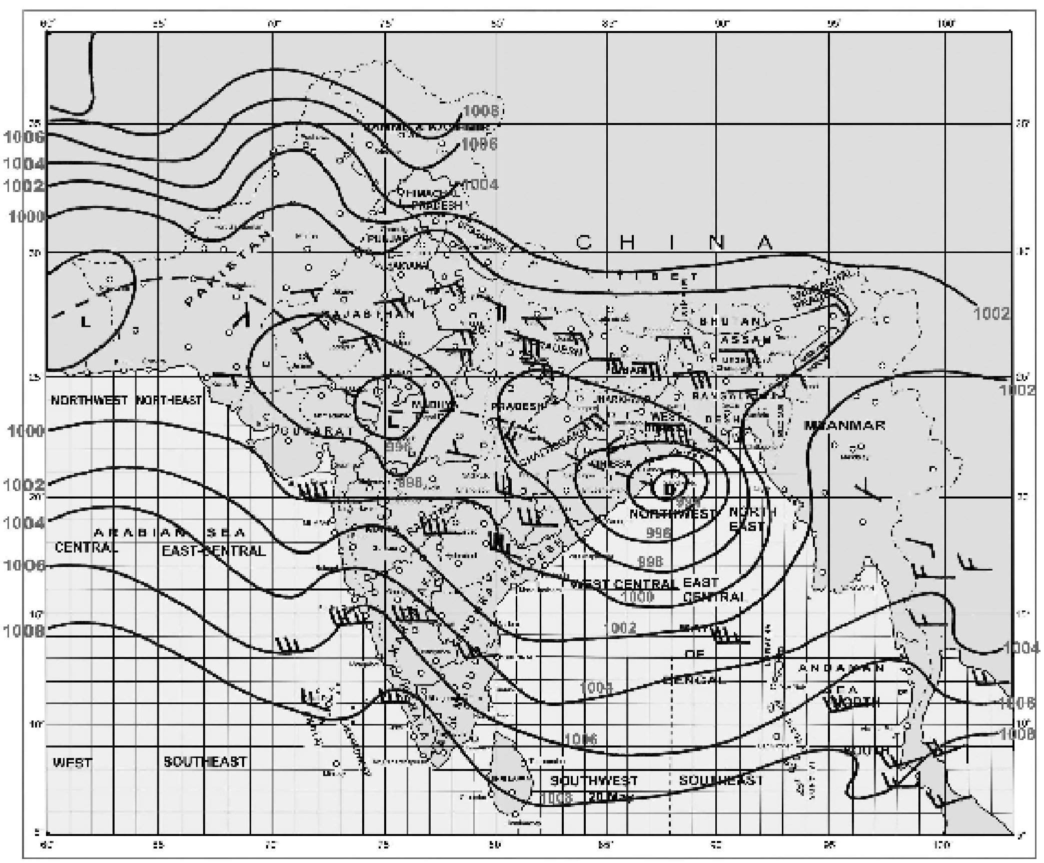


Figure 7a. Weather map for 16 August 2006. The contours of equal surface pressure (isobars) are marked. Note the depression caused off the coast of Orissa over the head of the Bay of Bengal. The symbol of a stick with teeth indicates the direction of the wind (wind at 1.5 km above the surface). The wind is along the direction of the stick blowing from the side where the teeth are. Wind speed increases with the number and size of the teeth.

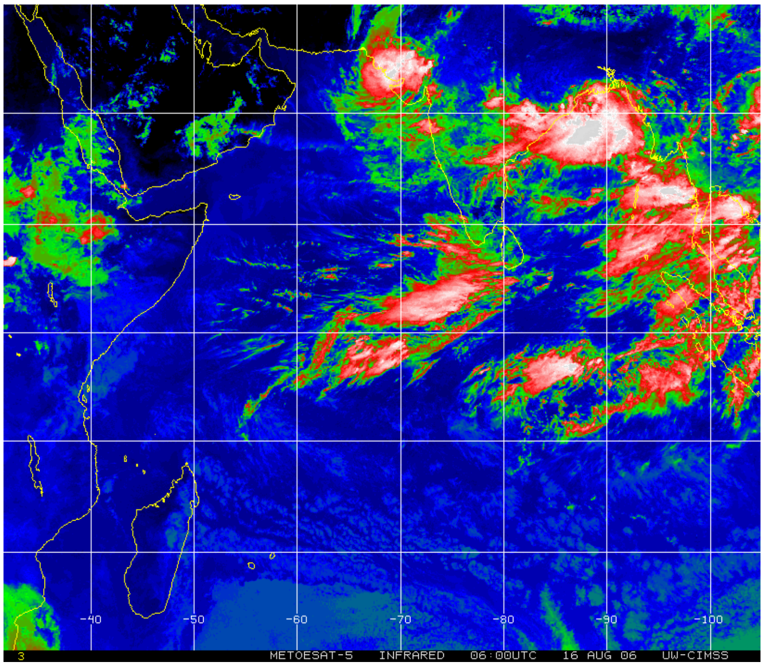
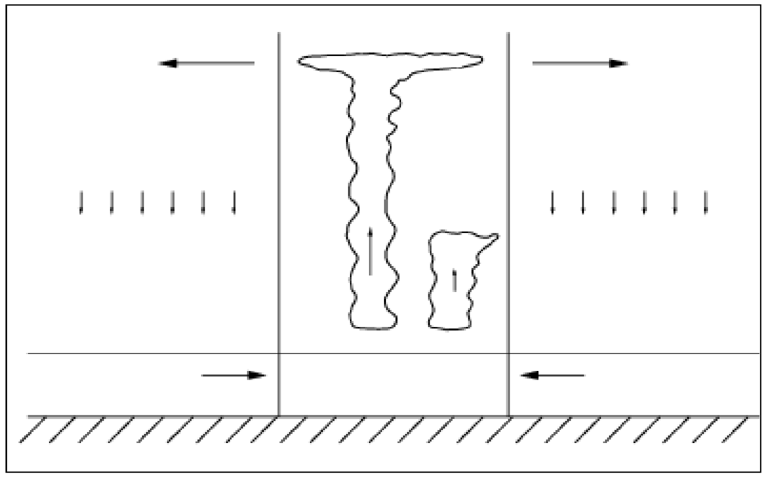


Figure 7b. Satellite picture for the same day corresponding to Figure 7a. Colours indicate the height of the radiating surfaces (cloud top, when there are clouds). White is for minimum temperature (i.e. maximum height of the cloud top), green for somewhat higher temperature (shallower clouds) and red is generally cloud free.

within the boundary layer friction also becomes important leading to a component of flow across the isobars. Convergence in the boundary layer is associated with this cross-isobar component.

The magnitude of the upward vertical velocity at the surface of the boundary layer, is proportional to that of the associated cyclonic vorticity above the boundary layer. When there is adequate moisture at the surface, this ascent can trigger genesis of cumulus clouds in the conditionally unstable tropical atmosphere. The convergence in the boundary layer and ascent of moist air occurs over regions of the scale of hundreds of km, and thus has hundreds of cumulus clouds embedded in it (*Figure* 8). The entire region gets heated because of the latent heat released in each such cloud. This leads to an intensification of the cyclonic vorticity which in turn leads to higher velocity of ascent. This positive feedback between the cumulus scale and the larger scales was first proposed by Charney and Eliassen [5] who suggested that the cumulus scale and larger scale motions support one another by this feedback, rather than competing. If

Figure 8.



there were a competition between these scales and the cumulus scale, the cumulus would always win. However, this cooperation between these two disparate scales, makes possible organization of clouds on these larger scales in the tropical atmosphere that we observe. This instability that leads to organized convection on synoptic and larger scales has been termed conditional instability of the second kind (CISK) to distinguish from the conditional instability that leads to the genesis of cumulus clouds. Subsequent to this seminal work by Charney and Eliassen [5], Gray [6], an expert on tropical cyclones, pointed out that the upward mass flux into the clouds is larger than convergence in the boundary layer. In fact these systems are characterized by convergence and ascent up to 3km or higher. Thus while convergence in the boundary layer is necessary, convergence in the lower troposphere is also required for intensification to the tropical cyclone stage.

Although the feedback proposed by Charney and Eliassen [5] is known to operate, it has been difficult to theoretically demonstrate that CISK can lead to a selection for convection on synoptic and larger scales. This was achieved by Srinivasan and Smith [7] who showed that in the presence of convergence in the boundary layer as well as in the lower troposphere there is selection for the larger spatial scales associated with tropical cyclones.

Concluding Remarks

We have seen that conditional instability of the tropical atmosphere plays an important role in all rain-giving systems ranging from an individual cloud (of spatial scale of about 5 km and temporal scale of an hour) to synoptic scale systems with time-scales of about 5 days and planetary scale systems with time-scales of about two weeks or longer. Another kind of instability – that of flow with vertical and horizontal shear in a rotating system, leads to the genesis of the synoptic scale vortex which intensifies to a depression or a hurricane. The critical role played by instabilities in the dynamics of the rain-giving systems, makes the problem of prediction inherently more difficult. However, the atmospheric models are able to generate reasonable predictions of how the synoptic scale systems will move, once they are generated. In the next article I will discuss the system responsible for the monsoon rainfall over the region.

Acknowledgements

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Suggested Reading

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