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ON

THE TROPOSPHERE, STRATOSPHERE AND MESOSPHERE

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NUMERICAL MODELS

Lectures 6 - 9

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NUMERICAL MODELS - P.K. Das

LECTURE VI

1. MONSOONS

Monsoons are periods of concentrated precipitation in the troposphere. They are largely confined to the tropics, and are caused by the different response of land and oceans to solar radiation. The thermal capacity of water is very large compared to air or land. Table 1 provides a few representative values.

Table 1

Thermal capacities (cal $cm^{-}S^{-1}$).

Substance	Thermal capacity
Air	3×10^{-4}
Water	1.0
Land	0.6

Apart from a wide difference in thermal capacity, it is to be noted that solar radiation is able to penetrate only a few centimetres of soil. But, because of the stirring action of wind blowing over water it is able to penetrate to much greater depths in oceans

Consequently, for an equivalent amount of solar radiation falling on soil, air and water, the rate of warming (or cooling) is the fastest for soil and air. Water, by comparison, warms (or cool) at a much slower rate.

This difference in the response of land and oce oceans is seen clearly in the day and night cycles of the sea breeze. During the day, the coastal areas warm at a faster rate than the adjoining sea. As the warm air rises over land, a cool see breeze sets in from the sea to land. At night, when solar isolation, is cut off the land cools much faster than the sea. Consequently, there is a breeze in the reverse direction, that is, from the land to the sea.

Monsoon winds are similar to sea breezes, but on a much larger scale. Similar to the day and night cycles of the sea breeze, we observe a summer and winter monsoon. The summer monsoon represents large scale flow of air from the oceans to the tropical areas of the Middle East, India, Southern Asia and West Africa. The winter monsoon is a similar flow of air, but in the reverse direction from land to sea from a belt of high pressure over Siberia to Southern Asia and Northwestern Australia. A wind reversal is also observed over West Africa. Over East Africa the asiatic summer monsoon coincides with a period of 'long' rains, while the

winter monsoon represents a spell of 'short rains'
The onset and duration of summer and winter monsoons are determined, approximately, by the seasonal movements of the sun.

As monsoons occur on much larger scales of motion, the movement of the air is deflected by the earth's rotation and by topographic barriers. Latent heat is released as water vapour condenses into liquid water during its traverse land.

Global monsoons are dominated by zonal and meridional cells, that is, cellular motion in the x-p or the y-p plane. These cells are referred to as Hadley or Walker cells. It has been conjectured that the variability of summer and winter monsoons, in terms of rainfall, is dependent on the relative intensity of Hadley and Walker cells. This needs verification with more quantitative data.

2. HEAT SOURCES AND SINKS

Much of the current research on monsoon is concerned with the location of heat sources and sinks in the atmosphere. This is of considerable interest because the intensity of differential heating provides the drive for monsoons. This is closely linked to the pattern of vertical motion in the atmosphere. As we have seen in earlier lectures, the mean vertical motion can be inferred from a combination of the

conservation of vorticity and the first law of thermodynamics. An early study by Das (1962) was concerned with mean monsoon conditions over India in the northern summer. By combining the equation for conservation of vorticity with the first law of thermodynamics, an equation was obtained for the height tendency of a constant pressure surface. The forcing function for this equation had a term for rate of non-adiabatic (known as diabatic in meteorology) heat supply, in addition to forcing by orography and friction. Computations were made to find out the diabatic heating needed to maintain a steady monsoon, that is, a circulation which will be independent of time. An interesting fact emerged: the radiation deficit over northwest India should produce an average cooling of 2.4° C (day^{-1}), while the latent heat released by excessive precipitation over northeast India should produce a warming of 3.2° C (day^{-1}) for a steady monsoon.

Large parts of northwest India are arid, where the soil has high reflectivity. Consequently, this is a region of radiation deficit, but the excess of outgoing terrestrial radiation over incoming solar radiation could only account for 50% of the required deficit. It was conjectured that scattering in an atmosphere with high dust load could account for the remaining part of the deficit. But, opinion is still divided on the role of scattering.

A few simple calculations indicate that the required warming over northeast India could be produced by the latent heat released by precipitation.

More recent evaluations of the heat balance have been made by Wei and Johnson (1982). The find an alignment of heat sources and sinks which shows movement northwards in conformity with the sun's northward shift as we approach the summer solstice. The alignment is, approximately, in line with the orientation of Walker and Hadley cells.

3. A VARIATIONAL APPROACH TO MONSOONS

Paltridge (1975) found that the broad features of the atmospheric general circulation could be reproduced, if we assume that atmospheric circulations minimize the rate of production of entropy in the atmosphere. What Paltridge did was to divide the earth's surface into 10 boxes, or latitude zones of equal area, and to compute the inter box flow of energy by minimising Q, where Q represents the ratio of the absorbed solar energy to the longwave terrestrial radiation radiated to space. In a sense, Q is the rate of entropy production summed up over the 10 boxes into which the globe was divided. Good agreement was observed between the predicted and observed meridional profiles of surface temperature, cloud cover and meridional energy flux by this model.

An extension of the minimization principle to simulate monsoon circulations has not yet been reported.

The conservation of vorticity (ζ) and the first law of thermodynamics are expressed by

$$\nabla^2\left(\frac{\partial\phi}{\partial t}\right) + f\vec{V}\cdot\nabla(\zeta+f) = f^2\frac{\partial\omega}{\partial p} \tag{3.1}$$

$$\frac{\partial^2\phi}{\partial p\,\partial t} + \vec{V}\cdot\nabla\left(\frac{\partial\phi}{\partial p}\right) + \sigma\omega = -\frac{kQ}{p} \tag{3.2}$$

where Q is the rate of diabatic heat per unit mass, f is the Coriolis parameter, ϕ represents geopotential (gz) and ω is the vertical velocity in pressure coordinates. The thermal stratification of the atmosphere is represented by

$$\sigma = -\frac{1}{\bar{\varrho}}\frac{\partial}{\partial p}\ln\bar{\theta} \tag{3.3}$$

where $\bar{\theta}$ is the mean potential temperature of the air and $\bar{\varrho}$ stands for density.

The thermal stratification is referred to as the static stability of the atmosphere. There are several ways of expressing this parameter. One could, for example, express it in terms of the Brunt. Vaisala frequency (N) by noting that

$$N^2 = g\frac{\partial}{\partial z}\ln\bar{\theta} \tag{3.4}$$

whence

$$\sigma = \left(\frac{RT}{g}\right)^2 \times \frac{N^2}{p^2} \tag{3.5}$$

Equations (3.1) and (3.2) may be combined to form a single equation for ($\frac{\partial\phi}{\partial t}$), which is the geopotential tendency. For convenience this will be referred to as ϕ_t.

We have

$$\left[\nabla^2 + \frac{\partial}{\partial p}\left(\frac{f^2}{\sigma}\frac{\partial}{\partial p}\right)\right]\phi_t = -f\vec{V}\cdot\nabla(\zeta+f) - \frac{\partial}{\partial p}\left(\frac{f^2}{\sigma}\vec{V}\cdot\nabla\frac{\partial\phi}{\partial p}\right) - \frac{\partial}{\partial p}\left(\frac{kf^2Q}{\sigma p}\right) \qquad (3.6)$$

where $k = R/c_p$, R is the gas constant and c_p is the specific heat at constant pressure.

Equation (3.6) is interesting because the forcing terms on the right represent (a) the advection of absolute vorticity $(\zeta+f)$ (b) the advection of temperature measured by $\frac{\partial\phi}{\partial p}$ and (c) the rate of non-adiabatic heating (Q) per unit mass. The left hand of the equation has an operator that is determined by the static stability (σ) . If the horizontal coordinates (ox, oy) are scaled by a characteristic length (L), and the vertical coordinate is scaled by (H) the scale height, we find

$$\nabla^2 \div \frac{\partial}{\partial p}\left(\frac{f^2}{\sigma}\frac{\partial}{\partial p}\right) \equiv \left(\frac{H}{L}\right)^2 \times \left(\frac{N}{f}\right)^2 \simeq 4 \qquad (3.7)$$

if a representative value of $f = 5 \times 10^{-5} S^{-1}$ is used and $N^2 \sim 10^{-4}\ s^{-2}$. We put $L \sim 10^6$m and $H \sim 10^4$m. But, as we can see, the ratio becomes very large if f

is small, as is the case near the tropics. This equation is consequently not valid in low latitudes.

The right hand side of (3.6) is dominated by the second term. For the third term to become co comparable with the second, very large rates of warming would be necessary. Such warming rates are only observed over limited regions of the earth's surface.

On the other hand, the advantage of (3.6) is that it permits us to compute the value of Q which will make ϕ_t vanish, that is, maintain a steady circulation.

The upper and lower boundary conditions are

$$\omega_1 = 0 \qquad (3.8)$$

$$\phi_t = (g\omega)_0 - \frac{1}{(\sigma\rho)_0}\left[\frac{kQ}{p} + \vec{V}\cdot\nabla\left(\frac{\partial\phi}{\partial p}\right)\right]_0 \qquad (3.9)$$

The subscripts 1 and o refer to the upper and lower boundary.

ϕ_t will vanish throughout the range of integration of (3.6), if the terms on the right of (3.6) and (3.9) also vanish. For a steady monsoon circulation this implies the constraints.

$$\frac{\partial}{\partial p}\left(\frac{kf^2Q}{\sigma p}\right) + f\vec{V}\cdot\nabla(\zeta+f) + \frac{\partial}{\partial p}\left(\frac{f^2}{\sigma}\vec{V}\cdot\nabla\frac{\partial\phi}{\partial p}\right) = 0 \qquad (3.10)$$

$$\frac{kQ_0}{p_0} = -\left[\vec{V}\cdot\nabla\frac{\partial\phi}{\partial p}\right]_0 + (g\sigma\rho\omega)_0 \qquad (3.11)$$

The vertical component of motion in (3.11), that is, ω_0 is in z-coordinates. It represents motion forced by orography and friction. The pattern of Q which satisfies (3.10) and (3.11) is then the one which will maintain a steady circulation, such as, the monsoon.

For global monsoons the converse is equally interesting. For example, given a pattern of Q what will be the best estimate of ϕ_t ? Recent advances in space technology have made it possible to determine patterns of Q on real time; consequently, the estimate of ϕ_t will have much value for prediction. This problem has not yet been solved.

4. TRANSPORT OF ENERGY

The sun's differential heating between low and high latitudes is balanced by the transport of energy by atmospheric waves, which are maintained by barotropic and baroclinic instability.

Over the years a very considerable volume of data have been compiled on transfer processes. To study the general circulation of the atmosphere, the motion is divided into a mean flow, and deviations from the mean. The latter are referred to as eddies. The mean represents an average both with respect to space and time. Four conversions are relevant:

(a) Differential warming to mean Available Potential Energy ($\bar{P}$).

(b) Available Potential Energy ($\bar{P}$) to mean Kinetic energy ($\bar{K}$)

(c) $\bar{P}$ to eddy available potential energy (P') and $\bar{K}$ to eddy Kinetic energy (K').

(d) Loss of Kinetic energy (K, K') by frictional dissipation and loss of eddy available potential energy (P') by warming or cooling on a small scale.

The main conversions in the tropsophere are from $\bar{P}$ to P', and from P' to K'. Figures for these conversions are available in standard texts (Holton, 1972); they support the view that the eddies generated by barotropic/baroclinic instability are primarily responsible for the poleward transport of heat to sustain the heat balance of the troposphere.

The energy transfer in the stratosphere is different from the troposphere because the high static stability of the lower stratosphere precludes large scale overturning by baroclinic instability. Both observations and theory suggest that the eddy Kinetic energy of the lower stratosphere is provided by the vertical propagation of eddy Kinetic energy from the troposphere.

There is also evidence to indicate that the polar stratosphere is warmer than the equatorial stratosphere. The mean temperature gradient is thus the reverse of what happens in the troposphere. Eddy motion in the stratosphere transports heat northwards, as in the troposphere; but, as the mean temperature gradient is now in the opposite direction, the transport of heat is up the temperature gradient. The primary conversion in the stratosphere is thus from eddy kinetic energy to eddy potential energy, because this is necessary to maintain the heat balance against radiative damping. This leads to a direct meridional cell with rising motion in the polar stratosphere and sinking air over the equatorial stratosphere.

Although the stratosphere is, comparatively speaking, less active than the troposphere, several interesting types of motion are observed there. Two interesting phenomena are spells of sudden warming in the winter polar stratosphere and the quasi-biennial oscillation of the equatorial stratosphere.

Sudden warmings are accompanied by a high degree of distortion of a vortex round the pole. This is followed by large scale warming of the polar stratosphere, which can reverse the prevailing temperature gradient. Warmings of as much as 40° C in a few days have been observed at 50 mb. Numerical models of sudden warming suggest that these sudden

spells of warming result from enhanced propagation of energy from the troposphere, especially by large tropospheric planetary waves. Sudden warmings are mainly observed in the northern hemisphere; consequently, it is believed that the tropospheric waves are forced by orography. The fact that these sudden warmings are only observed in some winters and not in others has not yet been satisfactorily explained.

5. FUTURE OUTLOOK

Numerical models have engendered a lively controversy among meteorologists. Some have even questioned their utility for the developing countries, especially in view of the large investment needed for computers.

Broadly speaking, they treat the atmosphere as a fluid for which the governing laws are known. Given the correct initial and boundary conditions, it will give an indication of the future state.

Another approach is based on satistical reasoning. It attempts to compute the future state by extrapolation of past history.

A new approach based on autoregressive models is beginning to emerge. This regards the past history of a meteorological variable as being made up of a component based on past history, and a component made

up of random numbers which best fit the time-series of data. The success of such models depends to a large extent on the discovery of a leading indicator. This is a feature which appears to modulate the time series. Numerical models help to locate a leading indicator. The sensitivity of atmospheric circulations to see surface temperature (SST), which was discovered by experiments with numerical models, is an example of a possible autoregressive model with SST as a leading indicator. An autoregressive model in India (Thapliyal, 1981) with the position of an upper air ridge acting as a leading indicator is another example.

The concept of predictability is engaging the attention of many scientists. Predictability is the upper limit of prediction by deterministic means. Lorenz (1969) showed that a certain amount of uncertainty is always present in meteorological data. When the growth rate of the fastest unstable modes is considered for example, the errors in initial data grow at the same speed as the unstable modes. By non linear interactions between different scale of motion the errors are spread to all the scales prevailing in the flow. By considering the predictability of different scales of motion, Lorenz (1969) observed that the theoretical upper limit of predicatability ranges from a few days to about 2 weeks.

It was suggested by Charney and Shukla (1981) that the time average features of tropical circulations, such as, the monsoons were more predictable than mid-latitude systems. But, for short range prediction the prospects were better in mid-latitudes because of the dominance of instability mechanisms. By way of comparison, numerical experiments by Shukla (1981) suggest that many features of time-averaged tropical circulations (such as monthly means) are the outcome of changes in boundary conditions at the earth's surface. Changes in albedo, and sea surface temperature are examples of physical processes that alter boundary conditions To what extent are year to year variations the outcome of fluctuations in boundary conditions remains a question to which the answer is not yet clear.

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LECTURE VII

1. MESOSCALE PHENOMENA - INTENSE ATMOSPHERIC VORTICES.

We have been concerned with atmospheric waves and vortices, whose linear dimensions are of the order 10^6m or more. There is another important class of meteorological systems, whose dimensions are smaller. But, despite their smaller dimension they are important because of their capability for generating large amounts of Kinetic energy. We refer to them as mesoscale features, or intense atmospheric vortices. A few examples are presented in Table 1.

Table 1.

Mesoscale Phenomena

Type of motion	Horizontal Scale (m).
Dust devils	1 - 10
Gusts	$10 - 10^2$
Tornadoes	10^2
Thunderclouds	10^3
Squall lines	$10^4 - 10^5$
Tropical cyclones (Hurricanes)	10^5

In view of their small scale, these systems are difficult to predict. The network of conventional land based data platforms is too coarse to capture the finer structure of these systems. Space based platforms (weather satellites) and improvements in radar technology have improved matters to some extent. The first two types in table 1 (dust devils and gusts) are of interest to commercial aircraft. Most national meteorological services are equipped with sensitive anemographs to record the frequency and intensity of gusts. Gusts are often the precussors of larger systems, such as, thunderclouds and squall lines The time scale of gusts a few minutes) is much smaller than the scale of the other types of motion in table 1.

2. TORNADOES

Tornadoes are intense vortices in the lower troposphere, which have a relatively short lifetime (a few hours) and small scale . This has hampered the collection of reliable data on their fine structure. Technological advances of the Doppler radar have now made it possible to monitor the growth of tornadoes associated with severe thunderstorms, but many aspects still remain unexplained. Simulation by high resolution numerical models and laboratory experiments form an important part of research on tornadoes.

Two features are required to produce an intense vortex; (a) a mechanism for producing rapidly rising mortion referred to as an updraft - in meteorology and (b) a source of angular momentum.

Ascending motion, it is believed, is associated with the buoyancy provided by the release of latent heat, or concentrated upward flux of sensible heat from the earth's surface. The latter is the cause of dust devils and, in some cases, tornadoes that are not associated with heavy rain.

It is still not very clear how the tornado acquires, and retains its angular momentum. The general view is that vorticity is generated by a baroclinic process, and this is intensified by convective activity. The production of vertical vorticity (ζ) is expressed by

$$\frac{D\zeta}{Dt} = \left(\frac{\partial u}{\partial z}\frac{\partial \omega}{\partial y} - \frac{\partial v}{\partial z}\frac{\partial \omega}{\partial x}\right) + \zeta\nabla.\vec{V} + \vec{F} \quad (2.1)$$

Here, (u, v, ω) are the components of motion in rectangular coordinates (ox oy oz), D/Dt is the total time derivative and $\vec{F}$ represents frictional dissipation. The first forcing term on the right of (2.1) is negligible for larger scale motions, but it is important for small scale phenomena. This term, in parentheses, contributes to vertical vorticity by tilting the horizontal vortex lines towards the vertical.

It is referred to as the 'tilting' term. On the other hand the second term on the right of (2.1) generates vorticity by stretching vertical vortex tubes. Consequently, we refer to it as the stretching term. The positive vorticity at the centre of a tornado is around $10^{-2}\ s^{-1}$. Of this amount the tilting term contributes $10^{-4}\ s^{-2}$, and the stretching term contributes $10^{-3}\ s^{-2}$. Clearly, the tilting term cannot be ignored.

Observations by Doppler radar indicate that tornadoes generate a hook-shaped echoe (Browning, 1964). The paucity of data makes it difficult to provide reliable information on the structure of a tornado, but it is characterised by a strong region of converging airflow and an updraft in regions near the ground, with an outflow and downdraft from regions aloft. The downdraft is believed to be the outcome of air being chilled by evaporation from rain (Fig. 1).

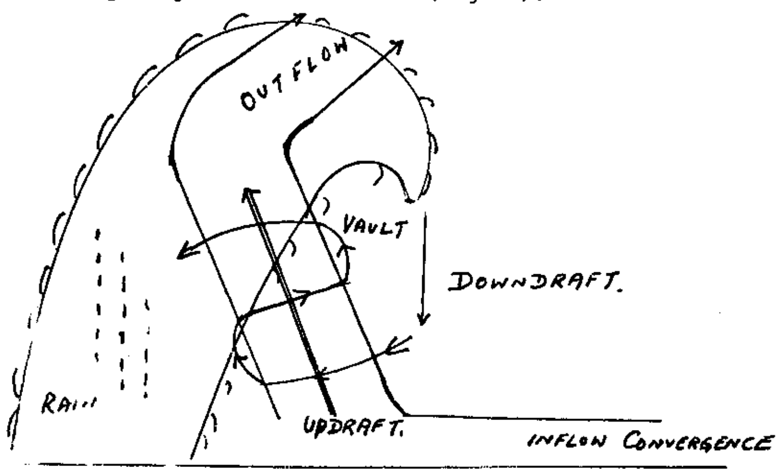

FIG. 1. (ADAPTED FROM BROWNING, 1964).

The downdraft eventually wraps itself round the central updraft. The clear region of the hook shaped echoe is called the vault. As the downdraft begins to circle round the central updraft, secondary tornadoes are often observed to develop at the interface between the updraft and the downdraft. A fairly comprehensive account of this process is provided in a recent article by Klemp and Rotunno (1982).

Numerical models have been developed in recent years to simulate the growth of tornadoes (Leslie and Smith, 1982). Very high resolution is needed for model simulation. The computational domain is about 8 km deep, and the resolution is about 25 m in the vertical and horizontal directions. Most models try to simulate a tornado by non-dimensional parameters. Four parameters are commonly used:

(a) Swirl parameter: (Average circulation round the outer edge of the convergence zone $(2\pi a v_\theta)$ ÷ Average volume flow rate $(2\pi a v_r)$)

(b) Radial Reynolds Number: $2\pi a v_r \div \nu$

(c) Internal aspect ratios: h/a

(d) External aspect ratios: r/a

where v_r, v_θ denote the radial and tangential velocity components, ν is the coeficient of Kinematic viscosity a stands for the radius of the central updraft funnel,

r is the radius of the outer convergence zone and h is the depth of the tornado.

A model developed by Snow (1982) was able to simulate the growth of subsidiary tornadoes or vortices when the swirl ratio was increased beyond a critical value. The maximum surface pressure deficit at the centre of a tornado occurred, however, when the tornado was a single vortex system.

Numerical simulation models are now being supplemented by laboratory models. It is interesting to form a review article by Maxworthy (1982) that one of the earliest experiments to simulate vortex formation was designed by Wilke in 1785. The experimental set up is usually made up of (a) a heated (or insulated) rotating/non-rotating base plate, (b) a rotating screen or angled vanes/jets, which form the side walls through which fluid may enter a rotating cylinder and (c) a suction pump at the top to simulate the vertical motion created by buoyancy. A reference to the several articles on laboratory models in a recent publication on Intense Atmospheric Vortices, edited by Bengtsson and Lighthill (1982), will provide the reader with fuller details. Laboratory models have succeeded in simulating the formation of secondary vortices, but it is still unclear to what extent the results are influenced by the geometry of the apparatus. Despite such

limitations, it is interesting to note the possibility of secondary vortices arising out of an instability mechanism operating on the downdraft-updraft interface of fig.1.

3. LINE SQUALLS

A phenomenon of considerable interest to Central and Western Africa and Northern India, is a band of cumulonimbus clouds, which are preceded by a line of altocumulus and stratus clouds. This is known as a line squall. On many satellite pictures line squalls are readily discerned by their shape and alignment.

In the tropics, squall lines tend to occur in regions where the wind speed has a maximum in the middle troposphere. This is typical of Central and Western Africa in the northern summer. In mid-latitudes, however, line squalls are associated with frontal discontinuities. Wind shears, generally undirectional, play a more important role in mid-latitude line squalls.

This feature has been simulated in a numerical model of African squall lines Bolton (1984). The model was linear, but for a profile which has a maximum in the middle troposphere, the model reproduces a succession of updrafts and downdrafts to the right of the main convection area. The general conclusion was that an appreciable amount of positive buoyancy was

necessary in addition to a wind maximum in the mid-troposphere.

Normally the temperature in the troposphere decreases at a rate of 5.5° C per km. But, as a rising parcel of air cools adiabatically at a much faster rate (10° C per km), it becomes cooler than its environment. Consequently, it tends to revert back to its original position. But, if the rising parcel is saturated, then further colling makes it shed its excess of water vapour by the release of latent heat of condensation. Its cooling rate is now much lesser than 10° C per km. The latent heat release could impart positive buoyancy to the rising parcel of air. This will enable it to continue to ascend until it is cooled by evaporation from falling rain, or eroded by dissipative forces. An atmosphere in which a rising parcel of moist air could become unstable through the release of latent heat is said to be conditionally unstable.

It was demonstrated by the late Professor J.G. Charney (1969) that a conditionally unstable atmosphere would favour convection on the scale of cumulus clouds rather than the larger scale meteorological waves. But, there need be no competition between the two because the meteorological wave, while feeding on the energy of cumulus clouds, could create conditions that favour the growth of cumulus clouds. A self sustaining mechanism

is thus set up which Charney and Eliassen (1964) named as onditional nstability of the S ond Kind (CISK).

Unfortunately, the concept of CISK has been a malter of some controversy over some aspects which are still unclear. Thus, Charney and Eliassen (1964) found that if a stably stratified atmosphere was disturbed, and the growth rate was computed for different wavelengths of perturbations, the doubling time of a perturbation of 100 km wavelength was about 1 day. Thereafter, the growth rate remained constant for all disturbances of smaller size, provided the atmosphere was unsaturated. To some extent, a similar result was obtained by Bolton (1984) who finds a maximum growth rate for a finite wave number, but Bolton's perturbations were sensitive to diffusive processes rather than the moisture content of the atmosphere.

4. TROPICAL CYCLONES.

Tropical cyclones represent another class of intense atmospheric vortices. They develop over tropical oceans in regions of warm sea surface temperature.

The balance of forces in a tropical cyclone is between the centrifugal force of winds rotating at speeds around 50-100 ms^{-1} and the pressure gradient force. The horizontal scale of a tropical cyclone is about 100 km. Two features distinguish tropical cyclones from tornadoes.

Motion on the scale of tropical cyclones remains in hydrostatic balance, but this is not the case for tornadoes or individual clouds. Contrary to tornadoes, there are large temperature fluctuations in tropical storms.

A tropical cyclone has a well defined clam central region. This is defined as the eye of the storm. This is not so well defined in tornadoes. The formation of the eye may be inferred by considering the angular momentum of an annular ring of rotating air

$$v_\theta r = \text{Constant} \qquad (4.1)$$

where v_θ is the tangential wind velocity and and r is the radial distance from the centre. As one proceeds from the outer periphery of the storm towards its interior (r decreases), the tangential velocity increases. But, the wind cannot increase indefinitely because the cyclone has a finite amount of kinetic energy. Unable to penetrate further, the wind tends to spiral upwards round a central region of calm subsiding motion. The radius of the eye is around 20 km, and the eye is warmer than its environment. Tropical cyclones are thus warm core systems.

What causes an incipient atmospheric vortex to intensify into a tropical cyclone is a matter of controversy. At one stage, it was believed that the genesis of a cyclone was associated with CISK through the

vertical motion generated in a boundary layer over the ocean surface. This is referred to as Ekman pumping of air. But, subsequent computations suggest that this was not sufficient to explain the rapid inflow of air needed to generate an intense vortex. The manner in which mesoscale systems, such as, cumulus towers are controlled by the larger scale wave motion in the tropics is not yet clearly understood.

Numerical models have been developed (Sundqvist, 1970) to simulate hurricane formation. A recent study by Pfeffer and Challa (1982) emphasizes the flux of momentum generated by eddy motion. Eddy motion here is represented by departures from symmetry around the axis of rotation. It thus appears that vortices that asymmetrical have greater potential for developing into mature cyclones.

Another model by Pearce (1982) emphasizes the role of buoyancy in the low level inflow of air. A distinction is made in this model between air that is channelled through clouds, and clearer air which does not directly come under the influence of moisture feed-back from the sea surface. This leads to the formation of secondary circulation at the interface between the inflowing air and the central calm region. The outflow in the upper troposphere, which is needed to balance the inflow of air in the lower troposphere, is expressed in a novel way with surface features.

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NUMERICAL MODELS - P.K. Das

LECTURE VIII

1. TROPICAL THUNDERSTORMS

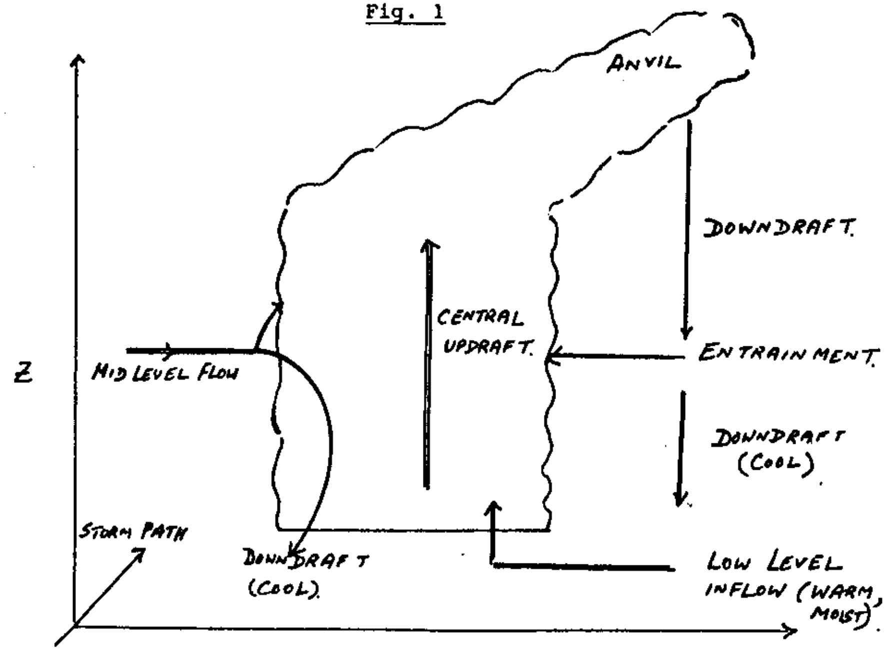
Tropical thunderstorms are well known for their destructive power. They represent a vortices which are not as large as a tropical cyclone and not as small as a tornado. Prominent tornadoes are often embedded in a thunderstorm. Such vortices are often referred to as "Supercells".

The structure of a thunderstorm has much similarity with that of a tornado, but on a larger scale. Recent observations with Doppler radar reveal a central updraught, which rotates cyclonically in the lower troposphere. This is replaced by divergent air flow aloft, which is anti-cyclonically curved. The upper flow is in the direction of the larger scale vertical wind shear; it is associated with an anvil shaped cloud. A downdraught is generated below the anvil by air that is cooled by evaporation from falling rain. The cool air spreads out laterally at the surface and interacts with the warmer moist air flowing into the thundercloud. As mentioned in the previous lecture, the interface between the downdraft and incoming moist air encourages the formation of new vortices, which are the supercells referred to earlier.

At the rear of the thunderstorm cell, a certain amount of mixing takes place between the environment and

the main body of the cyclone. This is known as entrainment. But, apart from entrainment, the environment tends to sink as if the thundercloud was a solid obstacle. The associated downdraft is again cool, and tends to rotate round the region of the central updraft. The structure is shown schematically in Fig.1.

Fig. 1



THUNDERSTORM CELL (ANTHES ET AL, 1978)

Many features of the thunderstorm have been now discovered as an outcome of improvements in radar technology. The principal ones are summarised below:

(a) The upper level divergence and the central updraft are more intense for severe thunderstorms.

(b) Hail is frequently observed near the cloud top, which is substantially colder than the ambient air.

(c) Severe thunderstorm developement appears to be preceded by a sudden increase in mid-tropospheric rate of ascent.

(d) Low cloud top temperatures observed in the "brightness temperatures" mapped by weather satellites appear to be associated with mature thunderstorms.

While these findings are still of a preliminary nature, improvements in satellite technology could well reveal more features in the next few years. The different stages in the life cycle of a thunderstorm beginning with its genesis, maturity and, finally, decay could well be revealed by changes in cloud top temperature.

A feature of the thunderstorm is strong interplay between its dynamics and thermodynamic aspects. By the latter we mean phase changes associated with the conversion of water vapour to liquid water (rain) and solid hail. These conversions are well understood, but it is not always easy to express them mathematically, because we do not understand the process of entrainment very well as yet. The updraught, for example, is controlled by entrainment; and this is equally true for the downdraught. In view of these difficulties, progress in modelling thunderstorms has been modest.

Srivastava (1967) was one of the earliest to design a simple one-dimensional model. The five uknowns

in his model were: the vertical velocity (u), temperature (T), the water vapour mixing ratio (q), the rainwater mixing ratio (r) and the mixing ratio for condenses water in clouds (W). The force of buoyancy may be represented by B, where

$$B = g\frac{(T_c - T_e)}{T_e} \qquad (1.1)$$

The subcripts c and e refer to the cloud and the ambient environment. The conservation equations for these five unknowns are

$$\mathcal{L}(U) = gB - g(w + r) \qquad (1.2)$$

$$\mathcal{L}(q) = E \qquad (1.3)$$

$$\mathcal{L}(r) = -\frac{1}{\rho}\frac{\partial}{\partial z}(\rho r v) - E_r \qquad (1.4)$$

$$\mathcal{L}(W) = E_c - P \qquad (1.5)$$

$$\mathcal{L}(T) + U\Gamma = -\frac{L}{C_p}E \qquad (1.6)$$

where $\mathcal{L} \equiv \partial/\partial t + U(z)\partial/\partial z$, and

and $g(W+r) =$ reduction in buoyancy due to weight of condensed water.

E = gain or loss of water vapour by evaporation or condensation.

E_c = cloud evaporation (or condensation).

P = rainfall from clouds.

E_r = evaporation from falling rain

ρ = density of air and

U = fall velocity of water drops.

The dry adiabatic lapse rate is represented by Γ.

These equations were solved by Srivastava for different initial conditions and assumptions. The model correctly simulated a gradual weakening of the updrafught, and an increase in the total water content of the cloud. In other words, the updraught velocity (U) and the total water content of the cloud are out of phase with each other. It is difficult to interpret this physically because the model has ∠entrainment built ∠no into it, but it shows that an increase in updraught does not immediately lead to an increase in water content.

More sophisticated models have been developed in the last few years to include more of cloud micro-physics and an extension to three dimensions. Murray and Koenig (1972), for example, employed a three dimensional model to infer that the cloud behaviour was strongly influenced by evaporation. This controls the central updraught, and generates a cold cloud top. The growth of this cold cap eventually stops the growth of the cloud, and generates a downdraught. Interestingly, the model postulates that the downdraught eventually stops inflow of air at lower levels, thereby leading to cloud decay.

The role of vertical shear in the development of a thunderstorm raises few questions to which the answers are not yet clear. As we have seen earlier, these shears generate vertical vorticity through the tilting of vortex tubes. This is most likely to be the source of vertical vorticity for the supercell tornado -like vortices, but one would expect that strong sheared flow would tend to inhibit the growth of a thunderstorm. The question has not been fully settled.

Prediction of thunderstorms is still a very difficult problem because of the explosive nature of convection in the atmosphere. Forecasters generally depend on data collected by upper air probes (radiosondes), and monitoring changes in wind direction and speed. The problem is still more difficult in regions with a sparse network of upper air stations. The principal features to look for are:

(a) a moist layer of warm air near to the surface.

(b) conditions which favour vertical overturning by adiabatic lifting of a layer of air so that it reaches saturation. This is known as potential instability in meteorology.

(c) a mechanism which could provide the ascent needed to release potential instability. This could be in the form of a wedge of warm moist air being undercut by colder air.

The forecasting skill by these methods is still limited, and the problem is made worse by the small lead time that is available for prediction in real time.

In an earlier lecture, mention was made of the fact that the source of energy for lower stratospheric circulations was the leakage of energy from the troposphere. In this context, the vertical propagation of gravity waves excited by tropospheric convection has been the subject of considerable research in recent years.

Clark and Morone (1981) find, for example, that intense heating of the mesosphere, which was measured by rockers launched from Wallops Island, Virginia, U.S.A. was accompanied by the passage of thunderstorms in the troposphere. They attribute the heating to the dissipation of gravity waves generated by thunderstorms. These waves may be a source of energy for the upper atmosphere in summer, when convection is most active. A study by Balachandran (1980) reveals gravity waves generated within the cold air outflow from thunderstorms. Gravity waves increase in amplitude at high altitudes because of the decrease in atmospheric density. Dopper VHF radar has been used to detect gravity waves in the upper troposphere.

Theoretical treatment of the interaction between cumulus clouds and gravity waves is difficult, because the evolution of both have the same time scale. As we have seen earlier, the parameterization of cumulus clouds

depends on their time scales being much smaller than the time scale of larger scale motion. But, for each wave we may define a critical level. This is the level at which the phase speed of the wave (c) becomes equal to the speed of the zonal current ($\bar{u}$) in which the wave is embedded. In an atmosphere with vertical shear there will be several critical levels for different wave numbers. It is yet unclear what happens at the critical level. Some have conjectured that the amplitude of the wave will be amplified at the critical level.

Observations by Doppler radar have provided new information on the characteristics of mesoscale turbulence. The analysis of energy spectra help us to identify the sources and sinks of atmospheric energy. If the spectral energy density depends on k^{-3}, where k is the wave number, it indicates that the motions are two dimensional because of the earth's rotation or strong thermal stratification. But, if the dependence follows a $k^{-5/3}$ law, it is indicative of three dimensional turbulence. The nature of the spectrum indicates whether energy is being transferred to larger or smaller scales of motion.

Observational evidence shows that the energy spectrum obeys the $k^{-5/3}$ law for wavelengths less than about 1000 km. This follows a well known law in the theory of turbulence which suggests a cascade of energy from larger to smaller scales by three dimensional turbulence. But, on the other hand, it has been argued

by Charney (1971) that atmospheric turbulence being geostrophic is two-dimensional. It should follow the $k^{-1/3}$ law. It is not yet clear whether the observed $-5/3$ law represents two dimensional turbulence in an inertial range which transfers energy from smaller to larger scales, as in mesoscale turbulence (Emanuel and Sanders, 1983).

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LECTURE IX

1. CHARGE SEPARATION IN A THUNDERSTORM

The electrical field in a thunderstorm resembles a vertical dipole with positive charge on the upper part of a thundercloud and negative charge near its base. The last four decades have seen several theories to explain the mechanics of charge separation. A comprehensive review of the earlier work has been provided by Mason (1971). In general, two classes of explanations appear.

The first, known as the precipitation hypothesis, emphasizes the tendency of larger hydrometeors to acquire negative charge. Their greater rate of descent compared to the smaller particles ensures an accumulation of negative charge in the lower parts of a thundercloud. The view has been expressed that the rate of charge separation by this process is not sufficient to explain the observed values of charge distribution in a mature thunderstorm.

The second hypothesis lays stress on convection as the main mechanism for the transport of electrical charge, and its accumulation in different regions of a thundercloud. Drake and Gunn (1946) observed charge separation during the melting of ice, and bursting of air bubbles released from ice during the melting process. Convection currents could influence charge production by this

mechanism. Many questions appear to remain unanswered on the relative merits of precipitation and convection for charge separation.

Vigorous vertical motion in a convective all often intensifies electrical activity. But, as we have seen earlier the intensity of an updraught and the quantity of water in a cloud are not always in phase, at least for one dimensional simple models of a thunderstorm.

2. RADAR REFLECTIVITY AND PRECIPITATION MECHANISM.

Doppler radar observations have shown that the presence of strong vertical motion (updraught), as revealed by intense radar reflectivity at altitudes around 6-7 km (-10 to -20° C), intensifies electrical activity (Lhermitte and Krehbiel, 1979). The vertical extent of the cloud, which is positively correlated with the strength of the updraught, is associated with lightning activity. Williams (1981) has reported that the rate of lightning flashes varies as the fifth power of cloud height.

The central updraught in a thundercloud has its origin below the freezing level of water. As it extends to a higher region of strong radar reflectivity, it is possible that supercooled water impinge on particles of graupel and form a coat of rime on them. The graupel - droplet collisions occur because of the difference in their terminal velocities.

As indicated earlier, Doppler radar observations also indicate a downdraught on the forward sector of the storm, which merges with the updraught in the region of high radar reflectivity. As the downdraft has its origin above the freezing level, it will contain ice crystals which will be carried to the updraught. This suggests ice-ice collisions as another mechanism for charge separation.

Experimental evidence (Mason, 1971) suggests that a growing layer of rime acquires negative electrical charge. The rate of charging depends on the rate of accretion. Computations by Mason (1971) suggest that the total charge generated by this process (around 500 c) is adequate to generate a lightning flash.

Mason has summarised the evidence for and against the separation of charge by ice-ice collisions; he does not support the idea of splintering ice crystals as an efficient mechanism. On the other hand, Lhermitte and Williams (1983) report that frequent electrical discharges occur 1-2 km above the high reflectivity region, with a much lesser number of cloud to ground flashes below. This suggests a more positive role for ice-ice collisions.

The role of different charged particles and the dipole nature of a thundercloud is not yet well understood, because the results of recent measurements appear to be inconclusive. Observations by Williams (1981) suggest

that the electrical and gravitational forces on hydrometeors (cloud drops) may be of comparable magnitude. Thus it appears that the terminal velocities of falling particles usually based on Stoke's law, may be appreciably modified, and earlier estimates of the efficiency of precipitation mechanisms need to be altered. The principal rain generating mechanisms - condensation on nuclei and collision leading to coalescence of drops - have been quantified without consideration of electrical forces, within a cloud. The interesting possibility arises of a cloud droplet rising under the action of a strong electric field, when it should be falling under gravity.

Lhermitte and Williams (1983) report an experiment in New Mexico with a vertically pointing Doppler radar to find the response of cloud droplets to an abrupt change in the electrical field caused by a nearby lightning discharge. A perceptible change in velocity of cloud drops was, however, noticed in only a few cases. It is not yet clear whether the field changes caused by lightning were, in fact, adequate to induce a change in fall velocity of drops that was larger than the noice level of a Doppler radar (about 1 ms^{-1}). This requires further investigation.

3. STRUCTURE OF LIGHTNING

The lightning discharge is the outcome of a build-up of charge in a thundercloud. Observations suggest that the ionization of the air may not be completely neutralised after the first discharge, so that the leader stroke of lightning is followed by multiple smaller strokes. The phenomenon is like the breakdown of the dielectric of a condenser when too great a field is put across the condenser.

In air, comparatively low potential gradients (200 volts per cm) are known to cause a corona discharge, They are seen to occur on radio antennas or elevated masts. The lightning discharge is associated with potential gradients as high as 30,000 volts per centimetre. Lightning does not occur until fields near 10,000 volts per centimetre are created.

Krehbiel (1981) and Latham (1981) suggest that the lightning charge favours an ice-ice mechanism rather than an ice-supercooled water mechanism for acquisition of negative charge in the region between -10° C to -20° C. But, it is not clear whether negative charge can not be transferred by lightning at altitudes less than -10° C level.

The nature of the electric charge transferred during lightning is inferred by simultaneous measurements of electric field changes at different locations during the same discharge. These observations suggest that the negative charge, associated with cloud to ground lightning,

is distributed horizontally and is confined to a narrow range of altitude and air temperature within the cloud. A close association is also inferred between the location of negative charge and echoes from radars directed upwards. It is not clear, however, whether the negative charge resides mainly on falling precipitation particles or on smaller cloud drops.

The detection of lightning from space by a satellite is beginning to provide new information on the frequency and distribution of lightning over the earth. Satellite observations indicate a significant difference between the lightning frequency over continents and oceans. The ratio is around 4:1. This observation is not surprising because of the greater cloud height over land.

An interesting observation is the observation of lightning in the atmospheres of Venus and Jupiter. The atmospheres of Venus and Jupiter are denser than the ea earth's atmosphere. Consequently, this probably implies greater electrical stress in these planetary atmospheres. Apart from density the chemical composition of the atmosphere over Venus and Jupiter. It is not yet clear how far these differences will have an impact on lightning in other planets but, clearly, an interesting area of future research is beginning to emerge.

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