(456)

SOME ASPECTS OF THE ENERGETICS OF THE ATMOSPHERIC GENERAL CIRCULATION

Dissertation submitted to the Jawaharlal Nehru University
in partial fulfilment of the requirements for the
award of the Degree of
MASTER OF PHILOSOPHY

C. SIVANESAN

SCHOOL OF ENVIRONMENTAL SCIENCES
JAWAHARLAL NEHRU UNIVERSITY
NEW DELHI-110067, INDIA
January **1983**

DEDICATED TO MY FAMILY

CERTIFICATE

The research work presented in this dissertation has been carried out in the School of Environmental Sciences, Jawaharlal Nehru University, New Delhi-110067. This work is original and has not been submitted in part or full for any other degree or diploma of any other university.

Sivanesan
C. SIVANESAN

G P Malik
G.P. MALIK
Supervisor

H K Das
H.K. DAS
Dean

DATE: 5. 1.83

SCHOOL OF ENVIRONMENTAL SCIENCES
JAWAHARLAL NEHRU UNIVERSITY
NEW DELHI - 110 067.

ACKNOWLEDGEMENT

I am grateful to Dr. H.S. Bedi, Director, India Meteorological Department, Lodi Road, New Delhi, who has sacrificed valuable time and effort to guide me at practically every step in the completion of this thesis. My thanks are due to my guide Dr. G.P. Malik, School of Environmental Sciences, J.N.U., for his keen interest and encouragement throughout this study. Special thanks go to my Senior, Dr. Suranjana Saha Attri for her constant help and supervison. I am also grateful to Dr. D.K. Bannerjee, Dean, School of Computer and Systems Sciences, J.N.U. and my friend Mr. Sunil Kumar Dhanawat for their valuable help in the computational work.

Thanks are due to Mr. Gajanan Hegde for typing the script and to Mr. Dipankar Sur for preparing the diagrams of the thesis.

Sivanesan

C. SIVANESAN.

CONTENTS

CHAPTER I

CONCEPTS OF THE GENERAL CIRCULATION OF THE ATMOSPHERE

1.1 INTRODUCTION

The general circulation of the atmosphere deals with the description and causes of large-scale flow of the atmosphere. To understand the wide range of physical processes that take place in our environment, we must first take a very broad view of our planet suspended in space. The following observations would comprise the salient points of such a view :

(i) The rays of the sun fall on the spherical surface of the earth with different angles of attack at different places; further, the angle of attack at any place changes as the earth turns on its axis.

(ii) There exists strangely irregular patterns of continents and ocean basins over the surface of the earth; the continents divide the world ocean into compartments;

(iii) There exists broad patterns of flow of air and water over the globe; of these, the atmospheric motion has an orderly belted pattern, whereas, the presence of continents is seen to inhibit the development of a uniform planetary system of water motion.

(iv) The atmosphere is seen to move at a high speed in contrast to the sluggish and often barely perceptible motion of the ocean waters.

Since the motion of water and air is a means of transferring energy from one region of the earth to another, the above observations might lead to the following conclusion: the continental surfaces, which are motionless, play the role of static receivers of energy from this motion, as also from the sun. It is the atmospheric motion that constitutes the dominant environmental control. Fig. 1.1 gives a system of latitudinal zones which is based on the latitudinal distribution of incoming solar energy.

As a first approximation, we now consider the circulation pattern that would be set up on an earth imagined not to rotate. Fig. 1.2(A) shows the setup as a spherical earth surrounded by a~~n~~ uniform atmospheric layer with no horizontal pressure gradient in existence. Imagine now that heat is applied to the equatorial belt, creating a net radiation surplus, while longwave radiation to space over the polar zones yields deficit and results in polar cooling. As shown, a pressure gradient from equator to poles has been setup

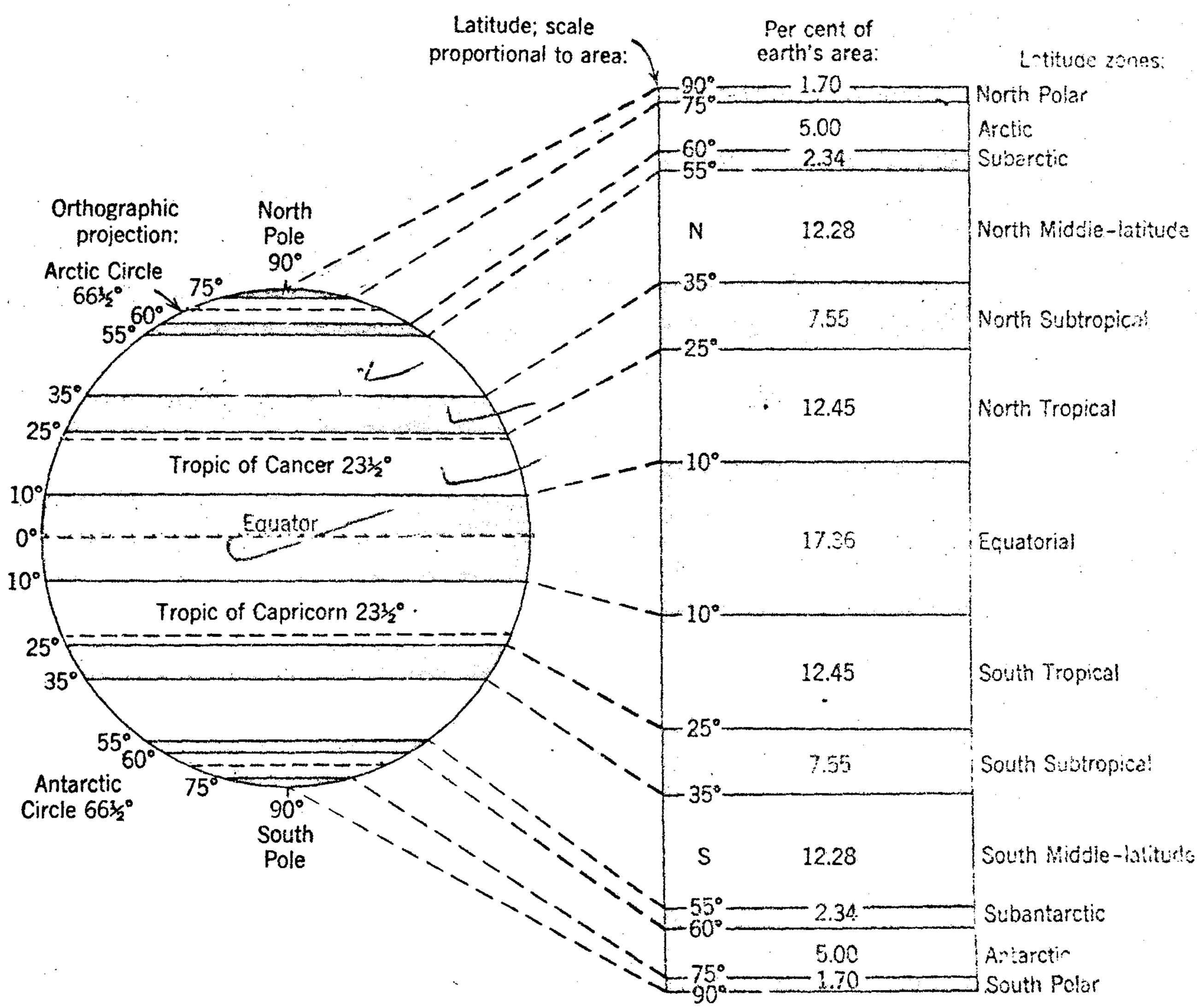


FIG. 1.1 **A geographical system of latitude zones.**

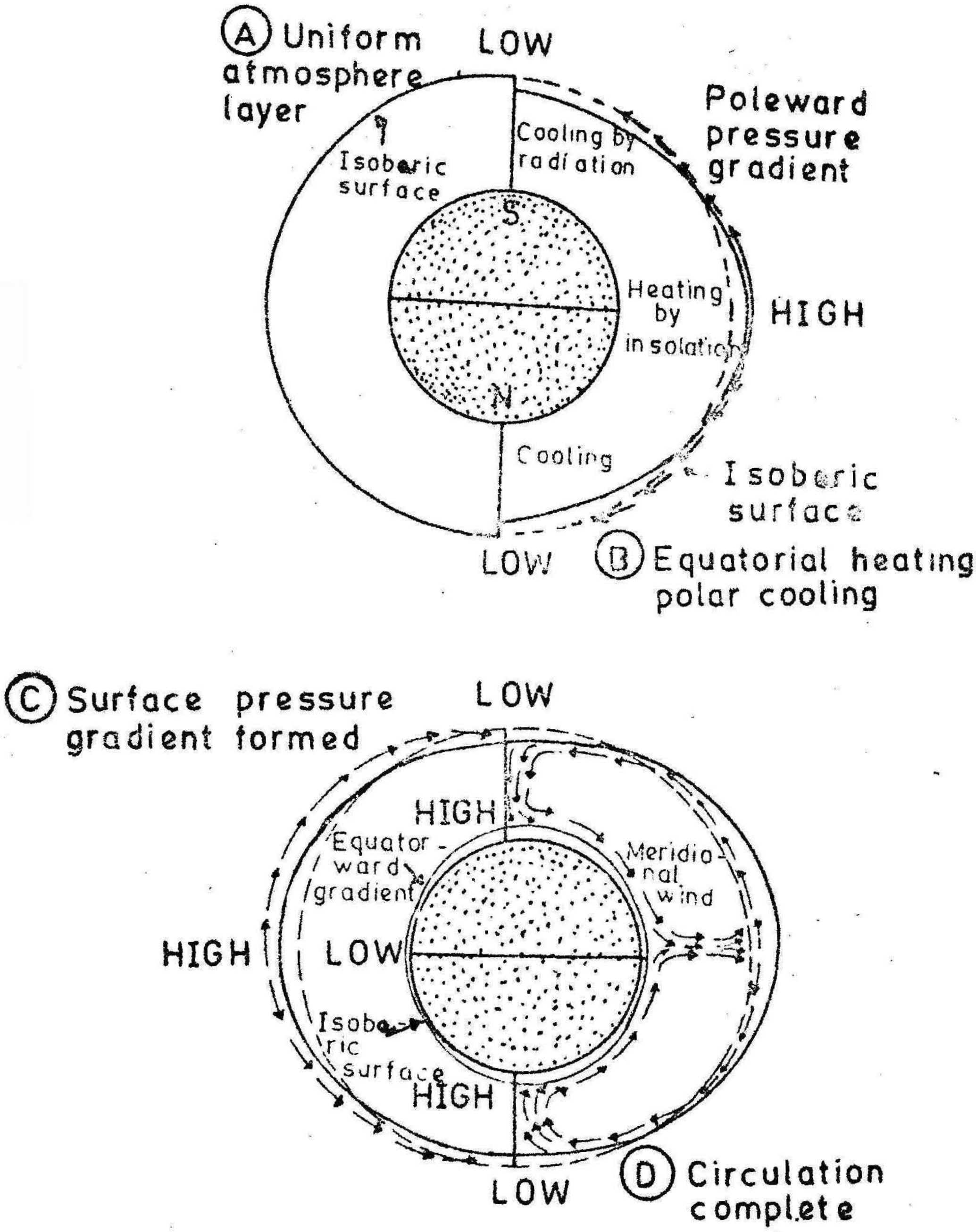


Fig.1.2 Convective wind system on an imagined non-rotating earth.

at high levels, causing poleward movement of air in Fig. 1.2(B). The redistribution of mass then causes formation of an equatorial belt of low pressure and two polar highs at low levels as in Fig. 1.2(C). Here the pressure gradient force is equatorward and causes surface winds to blow from the polar zones towards the equator. Circulation is completed by a rise of warmer air over the equatorial zone and a sinking of cooler air over the two polar zones. This simple model Fig. 1.2(D), although unworkable, does provide a first step in explaining the global circulation.

?

We now take into account the effects of earth's rotation by postulating an apparent deviating force (see Fig. 1.3) called the Coriolis Force, which except at the equator causes a particle to deflect to the right of its direction of motion in northern hemisphere, and to the left in the southern hemisphere. It is proportional to the velocity of the particle, the angular velocity of the earth about its axis and the sine of the latitude. A highly diagrammatic representation of the pressure and wind systems of the earth is shown in Fig. 1.4, as if no land areas existed to modify the belted arrangement of pressure zones.

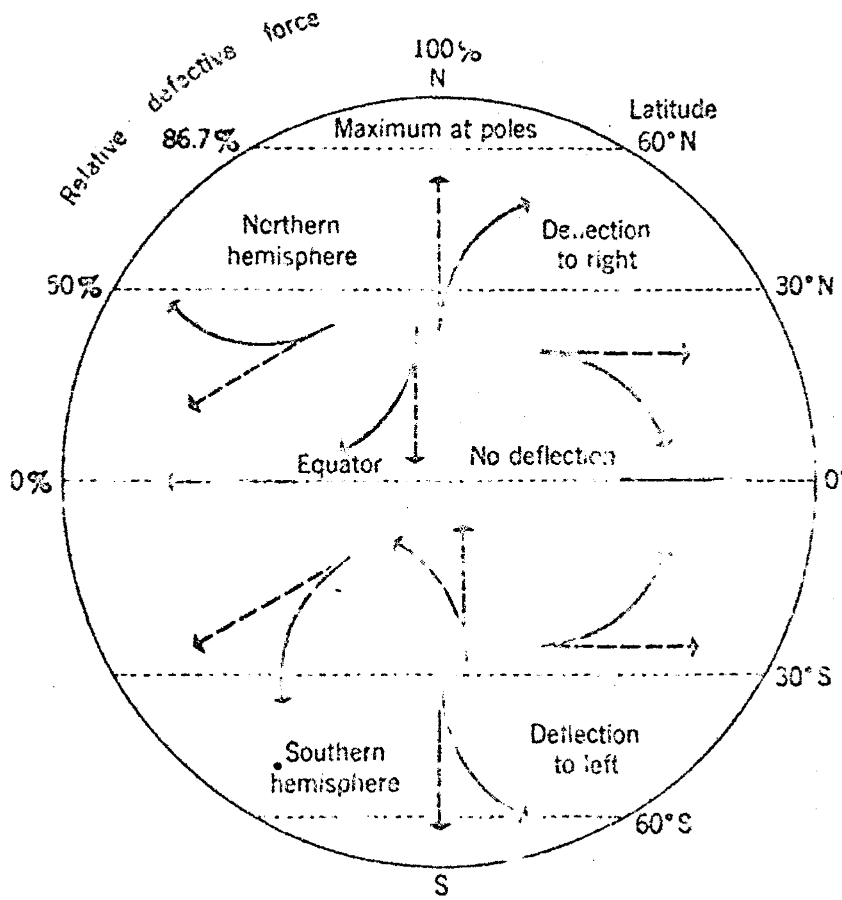


FIG. 1.3 **Deflective force of the earth's rotation.**(CORIOLIS FORCE).

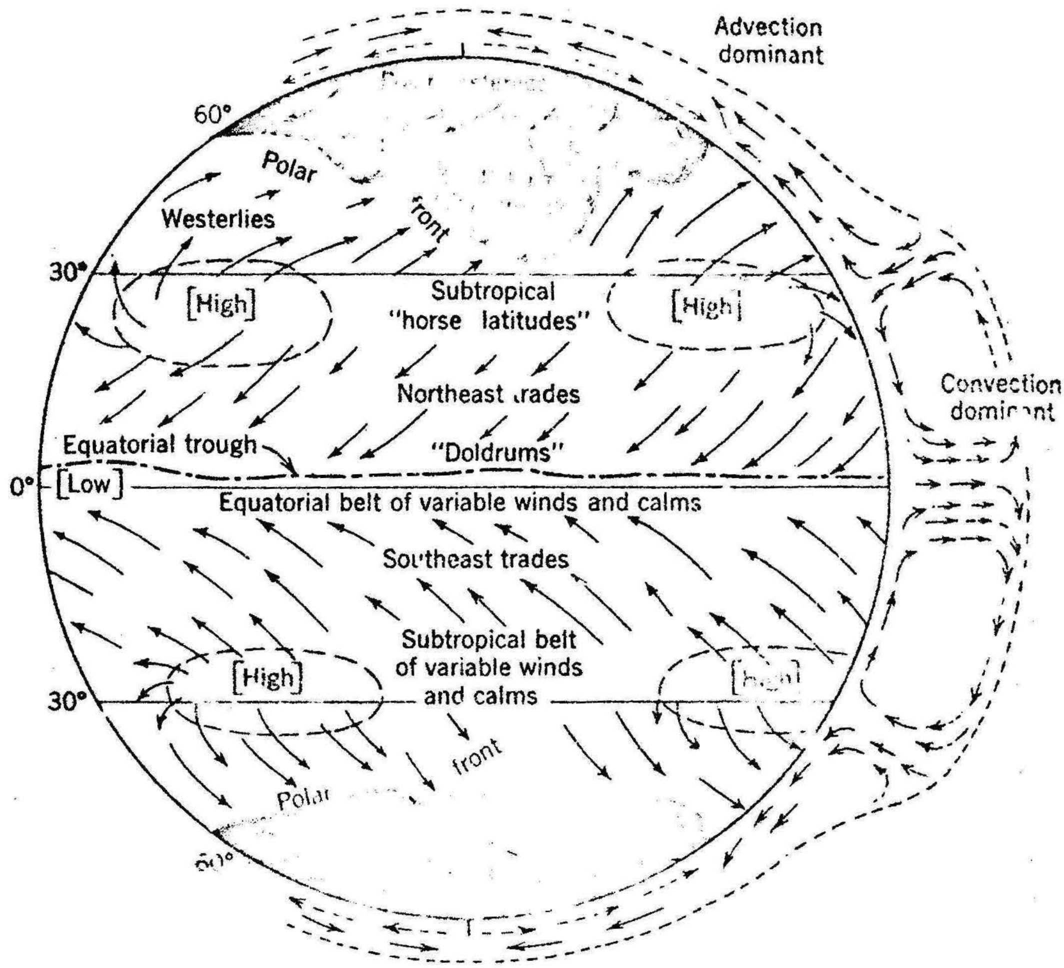


FIG. 1.4 The general scheme of atmospheric circulation on a rotating earth.

1.2 EARLY THEORIES OF THE GENERAL CIRCULATION OF ATMOSPHERE

The above picture of atmospheric circulation is a very idealized one. But the problem of explaining the circulation of even such an idealized atmosphere becomes difficult by the presence of advection, which is result of the displacement of the fields of motion and temperature by the field of motion itself. When averaged with respect to longitude, these advective processes appear to transport angular momentum and energy across the latitudes. Thus, the atmospheric advective processes carry these quantities towards the poles from the low latitudes and distribute them at high latitudes. This may be accomplished by a meridional circulation, i.e., a net equatorward flow at some levels accompanied and a net poleward flow at other levels. Thus a direct meridional cell would carry angular momentum and energy poleward, if it has equatorward flow in the lower levels and poleward flow at the higher levels.

The notable work of the astronomer Halley (1686) is the first detailed and methodical account of the trade winds as observed in three separate oceans. Seeking a common cause for them, he ascribed the north easterly trades in the northern hemisphere and south easterly trades in southern hemisphere to the tendency of the air to converge towards the most strongly heated region, i.e. the equatorial belt. Hadley (1735) further advanced Halley's idea. He noted that due to solar heating, air would rise in lower latitudes and sink at the higher latitudes, the circuit would then be completed with air moving towards the equator in the lower troposphere and towards the poles in higher troposphere. He also noted that in an absolute sense, the earth's surface moves most rapidly eastward at the lower latitudes and thus if air were moving initially towards the equator with no relative eastward or westward movement, it would try to conserve its absolute velocity and arrive at lower latitudes moving westward relative to the earth giving rise to the easterly trades. To explain the presence of the westerlies, he maintained that air moving directly poleward initially, would soon acquire eastward relative velocity to conserve its absolute velocity. On reaching the poles air would sink and become the prevailing westerlies (see Fig. 1.5).

The above theory requires some correction, because in the absence of eastward or westward forces,

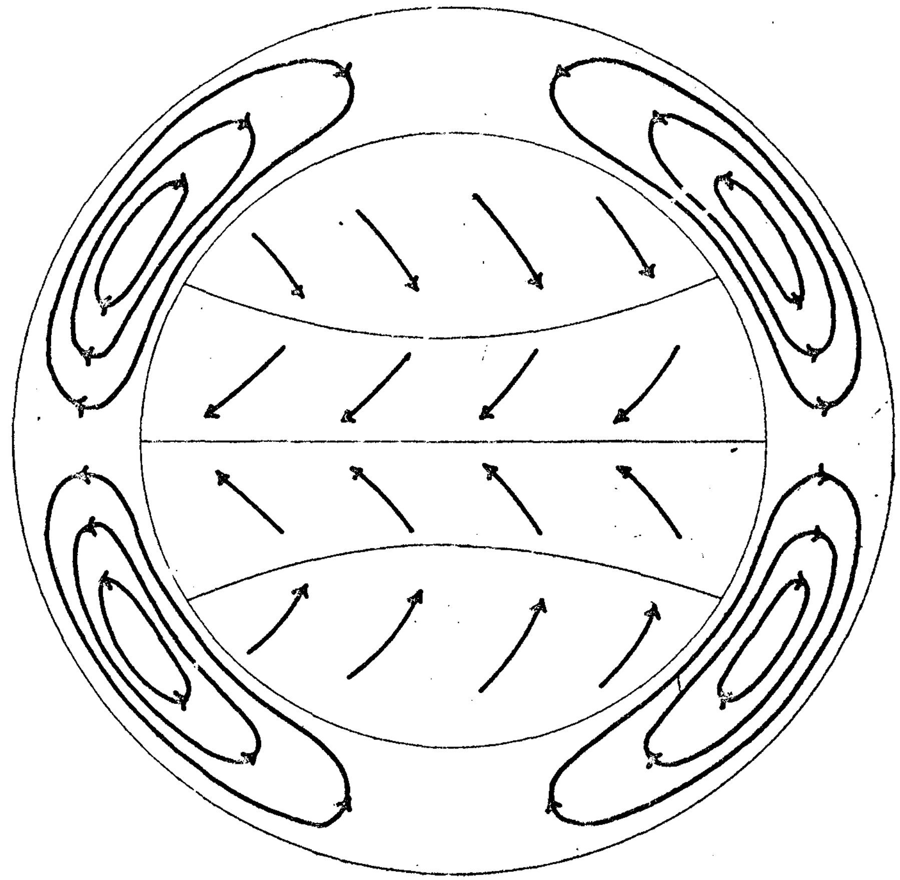


Figure 1.5 — A schematic representation of the general circulation of the atmosphere as envisioned by Hadley (1735)

air moving towards the equator or towards the poles conserves its absolute angular momentum rather than its absolute velocity. This fault arose, because Hadley had no knowledge about the north-south components of the Coriolis force. Also, he did not consider the thermal differential caused by the continents, which could have destroyed the symmetry of the heating or the presence of mountains and other barriers which could have distored the flow. Furthermore, he did not consider the presence of water vapour, whose thermodynamic properties were in any event not known in his day.

Dove (1835) accepted Hadley's theory only for low latitudes, but stated that the south-westerly winds in middle latitudes were a continuation of the south-westerlies above the trades, as their warmth and humidity demanded an equatorial origin. Thus, the trades would be a continuation of a return current from higher latitudes. He neglected the possibility that this current could occur in the upper troposphere as it appeared impossible for oppositely directed currents to cross in the horse latitudes without altering one another. He was thereby led to a scheme where south-westerly and north-easterly winds in middle latitudes flow side by side at different longitudes at the same level, rather than one above the other. The warm moist equatorial current was fed by the south-westerlies above the trades while the cold dry polar current fed the trade winds. He described the middle latitude storms as resulting from the conflict of two currents, i.e., the north and south currents.

Maury (1855) proposed a new general circulation theory, according to which, instead of a single meridional cell in either hemisphere, there are two cells - a direct cell like Hadley's within the tropics and an indirect cell in higher latitudes. The flow above the north-east trades is from the south-west and the upper level flow at higher latitudes is apparently supposed to be from the north-east. Maury was unable to give explanation for the cause of the indirect cells.

Ferrel (1856) was the first to explain that the normal pressure was not uniform over the Earth's surface. It was highest in the horse latitudes, and lower in doldrums, especially in the polar regions.

He stated that meridional currents from opposite directions do not cross but mix with each other and introduced a third cell in the general circulation theory. His great contribution was the introduction of a new force, the Coriolis force, appearing due to rotation of the earth which accounted for the previously unexplained features not only of the general circulation but also of cyclones and smaller disturbances. In this theory, he proposed:

1. The pole-to-equatorward density gradient, due to solar heating as proposed by Hadley leads to meridional motions, and hence, through the action of the east-west Coriolis force, to easterly and westerly winds.

2. Due to the action of the new force, the easterlies in low latitudes and westerlies in higher latitudes should be deflected away from the equator and the poles toward the subtropics, thereby creating the observed deficit of pressure at the equator and the poles, and the excess in the sub-tropics.

3. To explain the poleward drift in the surface westerlies, he observed that due to surface friction, the winds near the ground would be considerably

weaker than the winds somewhat higher up, while the horizontal pressure gradient would be reasonably uniform. The southward Coriolis force near the ground would therefore be insufficient to balance the pressure gradient and the westerlies would turn poleward later to rise and return equatorward.

4. Ferrel also noted that, for hydrostatic reasons, the latitude of highest pressure must be displaced toward the equator with elevation. He apparently felt that the opposing currents in the upper troposphere meet at this latitude in order to maintain the high pressure and thus inclined boundaries exist between low-latitude and middle latitude cells. In 1878, he further explored the possibility of deriving a mathematical expression for the general circulation (Ferral 1878), but was not successful as he was unable to formulate the frictional forces correctly. But this task was finally attempted by Oberbeck (1888) who represented the effects of friction by simple co-efficients of viscosity. He represented temperature by a simple analytical function of latitude and elevation and sought to derive the motion from the temperature field.

Thomson (1857) suggested that the low pressure at the poles resulted from the centrifugal force of westerly currents, which could be treated as large circumpolar vortices. Helmholtz (1888), deduced that, in the absence of friction, there should be easterly winds in low latitudes and westerly at higher latitudes. He then tried to find out how this circulation would change by the effects of heating and friction. Surface friction would produce a poleward drift in the surface westerlies. He maintained that the returning air above the trades must come into immediate contact with the cooler and more slowly moving air below, with the formation of a surface of discontinuity about vertical mixing. He thus concluded that the principal deterrent to stronger winds aloft was not surface friction, but the mixing of layers of different velocities by means of vortices forming on surfaces of discontinuity.

After Helmholtz, many scientists like Bort (1900) and Bigelow (1900) proved that only Ferrel's and Thomson's schemes were dynamically acceptable schemes. In 1921, Defant stated for the first time that the motion in middle latitudes was simply turbulence on a very large scale.

Another major contribution in the theories determining the role of cyclones in the general circulation was made by Jeffrey's (1926). He was concerned with the manner in which angular momentum was conveyed from low to high latitudes. He noted that

1) On large time-scales, angular momentum need not be considered, since its net transport is proportional to the net mass transport.

2) He stated that the net angular momentum transport was proportional to the product of the eastward and northward wind components.

3) Assuming a zonally symmetric flow with no meridional motion except within the lowest kilometers, or the friction layer, he found that the amount of angular momentum carried northward across middle latitudes was too small by a factor of at least 20 to balance the angular momentum transferred into the earth.

4) He concluded that the bulk of the required angular momentum transport was carried out by large-scale eddies, which he identified as cyclones, and pointed out that symmetric circulation is impossible i.e., the circulation must always be accompanied with eddies.

Bjerknes (1933) stated that a symmetric general circulation may be possible but would be unstable in some respects. In 1937, he concluded that the circulation which would prevail in the absence of departures from zonal symmetry was essentially the one given by Ferrel and Thomson, with a large direct cell occupying most of either hemispheres and a shallow indirect cell at low levels in middle latitudes. This circulation was unstable with respect to small zonally unsymmetric disturbances, hence the observed circulation would contain fully developed disturbances, which would assume the form of cyclones and anticyclones (see Fig. 1.7).

1.3 RECENT STUDIES CONCERNING THE GENERAL CIRCULATION OF THE ATMOSPHERE

As, towards the middle of twentieth century, the coverage of upper observations increased, many observational studies became possible to verify the earlier theories and also to postulate new concepts. Starr (1948) observed that the required northward transport of angular momentum could be produced by

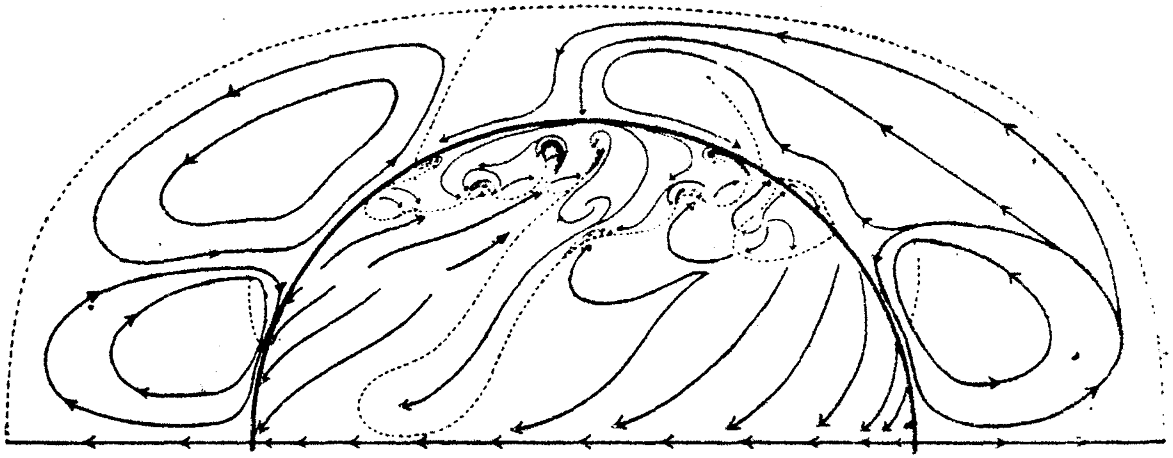


Figure 1.7.— A schematic representation of the general circulation of the atmosphere according to Bjerknes (1921)

troughs and ridges with a general NE-SW circulation. He stated that the shapes of the subtropical circulations are found to be such as to produce transport of angular momentum poleward. The downward flow of angular momentum in the westerly belts is effected by the presence of surface cyclones of the Bjerknes type in these regions. The upward flow in the easterly belts is effected through analogous mechanism, the details which were not clear owing to the lack of data. Bjerknes (1948) stated that elongated quasi-elliptical anti-cyclones with major axes oriented WSW-ENE could produce the same effect of transporting angular momentum poleward in the lower latitudes. Priestley (1948) observed that both the meridional circulation and eddies were important in effecting the required transports of angular momentum. He found no indication that the eddy transports were in agreement with mixing length concepts.

Convincing evidence of the importance of large scale eddies in the angular momentum balance was finally provided by Starr and White (1951). They found a large poleward transport by the eddies, but very small amounts of transport of angular momentum by the meridional circulation. In 1954, the same authors made a distinction between standing eddies and transient eddies. Van Mieghem (1952) and Lorenz (1955) calculated the kinetic energy of the mean motion, eddy motion, and also the Available Potential Energy of the mean and eddy fields.

Lorenz (1960) also deduced an energetically consistent set of atmospheric equations, both for the non-divergent and divergent flow. Wiin-Nielson (1962) separated the barotropic portion of the kinetic energy associated with the vertical shear. The work of Lorenz (1955, 1960), in particular, opened a new era in the general circulation studies considering various forms of energy in the atmosphere and their conversion and exchanges. Due to presence of large oceans data coverage over the tropical region and the southern hemisphere was rather scanty. The earlier studies on energetics of the atmosphere were mostly for the /et al. northern middle latitudes (Wiin-Nielson/1964, Saltzman 1961, Oort 1964). A comprehensive work on various aspects of general circulation of tropical region was undertaken at MIT, USA by Newell et al (1972). The studies on the atmospheric energetics, based on different data, from different sources and for different periods brought out some consistent features in energy processes of the atmosphere. Saltzman (1957) designed scheme for splitting up atmospheric flow into Fourier components and also transformed the equations of motion into their spectral (Fourier) form. This made possible the study of atmospheric energetics in terms of waves of different scales of motion. More details on the energetics of the atmosphere will be discussed in a subsequent chapter.

1.4 MODELS TO STUDY THE GENERAL CIRCULATION OF THE ATMOSPHERE

One of the recent and promising methods of studying the General Circulation lies in the development of "Mathematical Models" for the purposes of experiments, hypothesis-testing, and prediction. Thus the model is a mathematical formulation of the physical principles that govern the interactive processes which affect the behaviour of combined system of continents, oceans and atmosphere. The models could be either (i) physical models, or (ii) numerical models. In physical models, a liquid of known density and temperature profile is put in a rotating frame to approximately simulate earth's rotation and thermal field. Experiments with such physical models have provided interesting features of hypothetical atmospheres response to imposed thermal and rotatory forcings (Fultz, 1961).

In numerical models, the atmospheric physics is represented by a set of mathematical equations which can then be solved by modern computers to arrive at desired features of general circulation. As atmospheric equations of motion are highly non-linear, one has to resort to approximate numerical techniques for their solution. Owing to the tremendous range of scales of interacting atmospheric processes in relation to the limited spatial resolution of computational grids and data observing systems, it is not feasible to explicitly calculate the effects of small scale processes, in complete detail. Therefore, it is necessary to either relate the statistical effects of small scale processes, to measurable or explicitly computable conditions on larger scale, or to specify their effects more or less arbitrarily. The expression of the statistical effects of various small scale transport and transfer processes in terms of large scale variables explicitly resolved by the models is defined as parameterizations. Some common example of parameterizations are, the relation of net turbulent and/or convective transport of heat, momentum and moisture through the planetary boundary layer to conditions at the surface and at the top of the boundary layer. An explicit dynamic model of the atmosphere generally known as General Circulation Model can be used to study different features of the atmospheric general circulation. A first poineering attempt in this direction was made by Phillips (1963). A comparison of results based on actual data and those arrived at from general circulation models serve as a useful test on the validity of computational schemes and parameterization methods. Over last two decades efficient general circulation models have been evolved in different parts of the world, particularly in USA (NCAR GCM, i.e., National Center for Atmospheric Research General Circulation Models) and UK (The United Kingdom Meterological Office General Circulation Model), Geophysical Fluid Dynamics Laboratory (GFDL) Model which have been able to simulate various features of general circulation to a fairly good degree of accuracy.

1.5 SUMMARY

We have discussed in brief the evolution of research on the understanding of the atmospheric general circulation. Even though man has lived in the atmospheric environment from the beginning, his understanding of various aspects of the atmosphere which envelopes the earth became possible with the development of mathematical and physical sciences like physics, chemistry and geophysics. The early understanding was based on observation of the large scale flow of the lower atmosphere as observed by sea-men during their ocean voyages. Towards the end of seventeenth century and early eighteenth century, with the development in the physical sciences, the explanations of the observed features started emerging (Halley, 1686; Hadley 1735). These explanations, however, were not free from some fundamental discrepencies. Workers like Ferrel (1856), Helmholtz (1888), Jeffrey (1926) and Bjerknes (1935) further refined the concepts of the general circulation of the atmosphere. The availability of upper data over a larger part of the globe, particularly the northern hemisphere, provided opportunity for further refinment in the theoretical and observational studies of the general circulation of the atmosphere. Works of scientists like Lorenz

/et al.
/et al.

(1955, 1960), Wiin-Nielson/(1964), Saltzman (1957, 1961), and Newell/(1972) have put the understanding of general circulation of the atmosphere on a firm footing. The development of general circulation models of the atmosphere have provided further understanding to various observed features of the atmosphere.

The features like conservation of angular momentum, the transport of momentum and energy from lower to middle and upper latitudes, conversion and transformation of energy on various scales of motion and between mean flow and eddies are not fairly well known. However, as data with better resolution, particularly over tropics are becoming available, they will reveal more detailed features of various aspects of general circulation of the atmosphere. There is, therefore, an ample scope for further research on general circulation, based on both observational data and general circulation models for better understanding of the atmosphere, which may in the long run lead to its better prediction of atmospheric behaviour.

CHAPTER II

THE EQUATIONS FOR GENERAL CIRCULATION STUDIES

2.1 INTRODUCTION

The characteristic features of the circulation of earth's atmosphere vary in both space and time scales. The atmosphere is in an ever changing state resulting out of imbalances between different forces acting on it. However, there exist certain equilibria between some acting forces to good degree of approximation. The hydrostatic equilibrium defining a balance between force of gravity and vertical pressure gradient is a familiar property of large scale atmospheric flow. The balance between horizontal pressure gradient and Coriolis force known as gestrophic balance is another fundamental approximate property of the atmosphere which is valid mainly in the middle and higher latitudes. Such approximate properties of the atmosphere are useful in establishing physical relationship between different atmospheric variables in the course of observational or theoretical studies of the general circulation.

The general circulation of the atmosphere is the description of various statistics of the atmospheric flow based on sufficiently long data. The general circulation of the atmosphere is, therefore, derived from instantaneous atmospheric motion which is governed by the following fundamental laws of physics.

i) The law of conservation of momentum, or Newton's second law of motion.

ii) The law of conservation of energy or the first law of thermodynamics.

iii) The law of conservation of mass or the continuity equation.

2.2 BASIC EQUATIONS GOVERNING THE ATMOSPHERIC FLOW

2.2.1 EQUATIONS IN (x, y, z) CO-ORDINATE SYSTEM.

The above laws defining the large scale flows in (x, y, z) co-ordinate system can be expressed in terms of mathematical equations:

$$\frac{du}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv + F_x \qquad 2.1(a)$$

$$\frac{dv}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu + F_y \qquad 2.1(b)$$

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + F_z \qquad 2.1(c)$$

$$\frac{d\theta_T}{dt} = \frac{\theta_T}{C_p T}\dot{Q} \qquad 2.1(d)$$

$$\frac{d\rho}{dt} + \rho\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}\right) = 0 \qquad 2.1(e)$$

Equations 2.1(a) through 2.1(c) describe the principle of conservation of momentum in zonal, meridional and vertical directions respectively. The acceleration in horizontal components of momentum is the result of the horizontal pressure gradient force, Coriolis force arising due to earth's rotation, and frictional force, while the vertical acceleration of the momentum is caused by vertical pressure gradient, the force of gravity and frictional force.

Equations 2.1(d) governs the law of conservation of energy according to which, the change in potential temperature occurs due to external diabatic heat sources and sinks.

Equation 2.1(e) is the law of conservation of mass of a compressible fluid. According to this, the change in the density of air is caused by wind divergence.

2.2.2 HYDROSTATIC ASSUMPTION.

Large scale atmospheric flow is characterised by more or less horizonal motion, and the vertical component of wind or its acceleration is very small. As a result of this, an important approximation is the hydrostatic assumption mentioned earlier. An examination of comparative values of various terms in equation 2.1(c) shows that the terms other than term $\frac{1}{\rho}\frac{\partial p}{\partial z}$ and g are much smaller. Therefore, for large scale atmospheric flow, to high degree of accuracy, there exists a balance between the vertical pressure gradient force and force of gravity, or

$$\frac{\partial p}{\partial z} = -g\rho \qquad 2.2$$

This relation is known as hydrostatic balance.

2.2.3 EQUATIONS IN (x,y,p) CO-ORDINATE SYSTEM.

The hydrostatic approximation has great utility in studying the equations of atmospheric flow. Due to the diagnostic relationship defining the variation of pressure with height as indicated in equation 2.2, one can transform 2.1(a) through 2.1(e) from the (x,y,z) co-ordinate system to (x,y,p) co-ordinate system. When this is done, the equation 2.1(e) reduces to:

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} = 0$$

where $\omega = \frac{dp}{dt}$ is the vertical velocity in p system. As an air parcel moves vertically relative to pressure surfaces, the rate of change in its pressure is a measure of vertical velocity. This form of continuity equation has advantage that the density variation with time does not appear in the equation explicitly. A compressible fluid like air in (x,y,p) system can, therefore, be treated like an incompressible fluid without any sacrifice in its physics.

The pressure gradient terms $\frac{1}{\rho}\frac{\partial p}{\partial x}$ and $\frac{1}{\rho}\frac{\partial p}{\partial y}$ in (x,y,p) system take the form $\frac{\partial \phi}{\partial x}$ and $\frac{\partial \phi}{\partial y}$.

where, $\phi = gz$ is the geopotential. Thus, in the co-ordinate system, the density of the air does not appear explicitly in the horizontal momentum equation also.

In (x,y,p) system the above set of equations along with hydrostatic assumption may be written as

$$\frac{du}{dt} = -\frac{\partial \phi}{\partial x} + fv + F_x \qquad 2.3(a)$$

$$\frac{dv}{dt} = -\frac{\partial \phi}{\partial y} - fu + F_y \qquad 2.3(b)$$

$$\frac{dp}{dz} = -g\rho \qquad 2.3(c)$$

$$\frac{d\theta_T}{dt} = \frac{\theta_T}{c_p T}\dot{Q} \qquad 2.3(d)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} = 0 \qquad 2.3(e)$$

and if we add the equation of state

$$\frac{p}{\rho} = RT \qquad 2.3(f)$$

this becomes a close set of six equations in six variables, u, v, ω , p, ρ and T, or θ_T , provided that the frictional force, F_x and F_y and diabatic heat source or sink $\dot{Q}$ are known or are expressed in terms of the above six known variables. The process of expressing physical processes like frictional effects, diabatic heating or cooling, convective processes in terms of measurable meteorological parameters like temperature, wind components, etc. is called parameterization.

For many purposes it is not necessary to study or solve these equations in their above form. Often approximate equations retaining the essential properties of the original equations are of much practical utility. The method of scale analysis based on an examination of comparative magnitude of different terms of the equation of motion is useful for arriving at such consistent approximations. One of the very useful approximation is that of geostrophic approximation.

2.2.4 GEOSTROPHIC ASSUMPTION.

The scale analysis of horizontal momentum equation 2.3 (a) and 2.3(b) shows that atleast in the middle and higher latitudes the pressure gradient term $\frac{\partial\phi}{\partial x}$, $\frac{\partial\phi}{\partial y}$ and fv, fu are of a comparable magnitude and atleast an order larger than the other terms. Hence there exists a balance between the pressure gradient force and Coriolis force to a good degree of approximation in the middle and higher latitudes. Thus, for many purposes these two equations may be written as

$$fu = -\frac{\partial\phi}{\partial y} \qquad 2.4(a)$$

$$fv = \frac{\partial\phi}{\partial x} \qquad 2.4(b)$$

This relation between pressure gradient force and Coriolis force is known as geostrophic assumption and is valid in middle and higher latitudes. However, in the equatorial region where the value of f becomes very small, or in the atmospheric boundary layer where the frictional forces dominate, geostrophic approximation is not valid.

2.2.5 **THERMAL WIND EQUATION.**

Another useful relation is the relation between the horizontal temperature gradient and the vertical shear of the wind, known as the thermal wind equation. This can be derived by differentiating equations 2.4(a) and 2.4(b) with respect to 'p' and making use of equation 2.3(f).

The thermal wind equation derived thus, has the form

$$\frac{\partial u}{\partial p} = \frac{R}{fp}\frac{\partial T}{\partial y} \qquad 2.5(a)$$

$$\frac{\partial v}{\partial p} = -\frac{R}{fp}\frac{\partial T}{\partial x} \qquad 2.5(b)$$

This equation is useful in calculating the vertical wind shear from horizontal pressure gradient or vice versa.

2.3 DECOMPOSITION OF GENERAL CIRCULATION STATISTICS INTO MEAN AND EDDIES

The atmospheric circulation exhibits variations ranging over different scales, both in space and time. In the time domain, besides the changes on seasonal and longer scales, the atmosphere poses many variations of shorter durations. There are also varied scales of motion in space ranging from planetary scale systems to cyclones of relatively much smaller horizontal extent. Thus, for a better understanding of general circulation of the atmosphere in terms of scale interactions of different waves, it is advantageous to divide the atmospheric flow in terms of basic flow and the eddies. An observed feature of the atmospheric circulation is that the wind blows dominantly along the latitude circles. Thus, the basic flow is taken as the zonally averaged flow along latitude circles, so that along any latitude an atmospheric variable ϕ may be assumed to consist of two components $\bar{\phi}$ and ϕ' such that

$$\phi = \bar{\phi} + \phi' \tag{2.6}$$

where $\overline{\phi} = \frac{1}{2\pi}\int_0^{2\pi} \phi \, d\lambda$ is the zonal average of ϕ

and ϕ' = its local deviation from the zonal average.

A similar process can be adopted in time domain also. If we split the atmospheric variable ϕ at any time into two components, i.e., the time average and deviation from it,

we may write

$$\phi = [\phi] + \phi^{*} \tag{2.7}$$

Where, $[\phi] = \frac{1}{2T}\int_{-T}^{+T} \phi \, dt$, is long term average over time interval 2T, and

ϕ^{*} = the deviation of instantaneous value from this average.

Combining (2.6) and (2.7) we decompose the atmospheric variable ϕ as

$$\phi = [\overline{\phi}] + \overline{\phi}^{*} + [\phi'] + \phi'^{*} \tag{2.8}$$

where, $[\overline{\phi}]$ represents time averaged zonal mean field

ϕ^* represents temporal variation of zonally averaged field,

$[\phi']$ represents standing eddies, and

$\phi^{*\prime}$ represents transient eddies.

Using the same decomposition scheme the product of two variables (as an example u and v components of wind) may be expressed as

$$u.v = \left([\bar{u}] + \bar{u}^* + [u'] + u'^*\right).\left([\bar{v}] + \bar{v}^* + [v'] + v'^*\right)$$

and when both sides are averaged over space and time, we get (Starr and White, 1952)

$$[\overline{uv}] = [\bar{u}][\bar{v}] + \overline{[u'][v']} + [\overline{u'v'}] \qquad 2.9$$

Left hand side of (2.9) gives the zonally and time averaged meridional flux of westerly momentum, which is contributed by

i) The mean meridional motion - $[\bar{u}][\bar{v}]$

ii) Standing eddies - $[\bar{u}'][\bar{v}']$ and

iii) Transient eddies - $[\overline{u'v'}]$.

Thus, decomposition of meterological fields into mean and eddy components, can bring out many physical processes prevailing in the atmospheric general circulation.

One may further decompose the eddy motion into Fourier components and also transform the equations of motion in spectral or Fourier domain (Saltzman, 1958). This procedure enables one to study the general circulation of the atmosphere in terms of different scales of motion. Further, details on such decomposition will be discussed in a subsequent chapter.

2.4 SUMMARY

Most of the General Circulation studies are based on the set of equations 2.3(a) through 2.3(f), after decomposing the meteorological fields into mean eddies as in equation 2.8 and 2.9. A number of observational studies have been made on these lines using the global observations (White, 1951; Starr 1951; Oort and Rasmusson, 1971; Newell et al., 1972). In the following chapters we will describe the various features of the general circulation of the atmosphere as derived from the analysis of these equations and decomposition schemes. It may, however be, not possible to give full description of the complete analysis, unless it is necessary.

CHAPTER III

OBSERVED FEATURES OF ATMOSPHERIC GENERAL CIRCULATION

3.1 INTRODUCTION

The characteristic features of general circulation of earth's atmosphere have got many heterogeneties, due to inherent instability in the atmospheric flow system, the distribution of heat sources and sinks and geographical distribution of oceans and continents. If the atmosphere were completely steady, the task of studying its general circulation could be a straight forward matter of geographical exploration. If the circulation patterns varied only with time of the day or the season of the year, but not otherwise, the task of general circulation studies would be prolonged, but will not be still too complicated. But the fluctuations in the atmospheric circulations occur on all time and space scales to include simple gust and lulls in local winds to the development and decay of individual thunderstorms, the passage of migratory cyclones and anti-cyclones, the extended period oscillations in zonal index and changes corresponding to global atmospheric system, like monsoons.

It is, therefore, difficult to describe all features of atmospheric circulation to finer details. The general circulation studies, therefore, concentrate on the broad description of significant atmospheric features and their variations with season or periods of time, as well as their possible causes. In this chapter we will be concerned with such broad features of global atmospheric circulation and their physical causes, as we understand them today.

3.2 LONG TERM ZONALLY AVERAGED WIND AND TEMPERATURE PATTERNS

If we move along a latitude circle we encounter wavelike eddies of various scales. The meridional wind from northerly and southerly directions associated with these eddies largely compensate each other, so that, the zonally averaged wind becomes practically zero. The zonally averaged wind field is, therefore, described essentially by the zonally averaged zonal (east-west) component of the wind and will be of interest for our study.

3.2.1 ZONAL WIND:

Fig. 3.1(a) and 3.1(b) show the long term averaged of zonally averaged zonal component (u) of wind extending from north pole to south pole during the northern winter and summer seasons, based on the studies of Buch (1954) and Obasi (1963). The following features are evident from these figures.

i) The prevalence of easterlies over the tropical latitudes and westerlies over the middle and high latitudes.

ii) Existence of westerly wind maxima in both hemispheres in the upper troposphere during both seasons.

At lower levels the easterlies extend between about 30°N and 30°S. The latitudinal extent of easterlies shrinks with height and is only from 10°N to 10°S in the upper troposphere during summer, while during winter the easterlies in the upper troposphere do not exist at all. During winter the strongest easterlies occur in the northern hemisphere along about 10°N in the lower troposphere. During summer season easterly maximum occurs in the upper troposphere near the equator corresponding to the location of easterly jet over India during monsoon season.

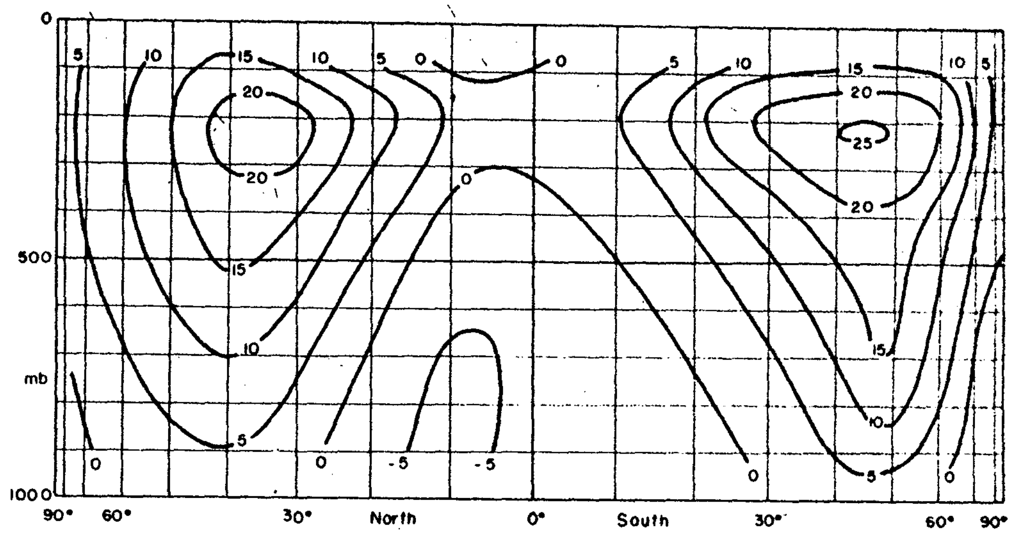


Figure 3.1 (a) The time-and-longitude averaged zonal wind $[\bar{u}]$ in northern winter and southern summer (October-March) as estimated by Buch (1954) and Obasi (1963). Values are in m sec^{-1}

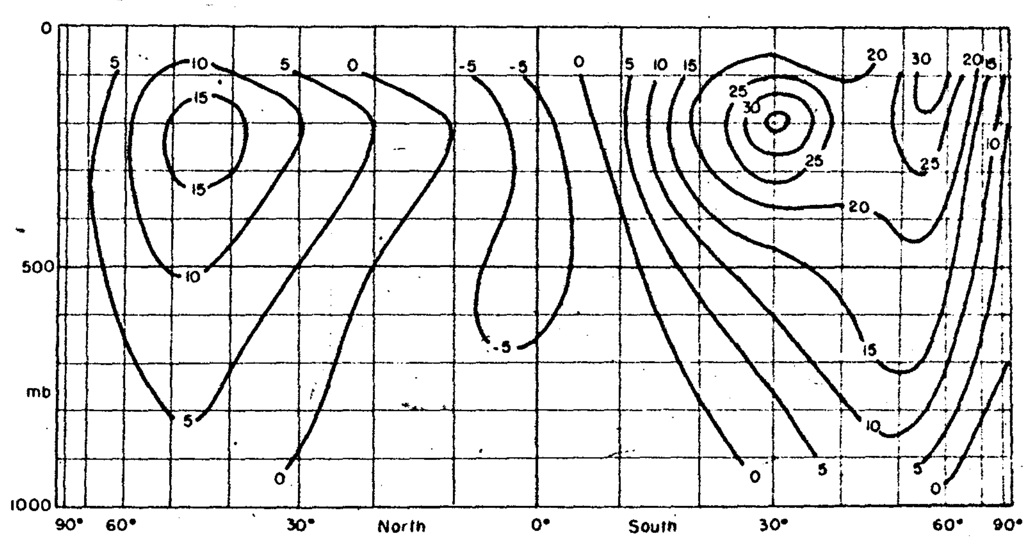


Figure 3.1 (b) The time-and-longitude averaged zonal wind $[\bar{u}]$ in northern summer and southern winter (April-September) as estimated by Buch (1954) and Obasi (1963). Values are in m sec^{-1}

The westerlies in both the hemisphere gave greater latitudinal extent in both hemisphere and have maximum strength near 200 mb during both the seasons. These wind maxima occur over the region where westerly jet streams blow in the two hemispheres. During the winter seasons of each hemisphere 10°-15° and lies at about 45° latitude. The intensity of westerly wind maxima is more during winter than during summer, also the southern hemispheric westerly maxima are stronger than northern hemispheric maxima during respective winter and summer seasons.

3.2.2 <u>TEMPERATURE</u>

Fig. 3.2(a) and 3.2(b) shows the zonally averaged long time mean temperature during winter and summer seasons after Palmen and Newton (1867). During both the seasons the temperature is maximum over equatorial region and decreases towards poles. The equator to pole temperature gradients are weak over equatorial region and became very strong in middle

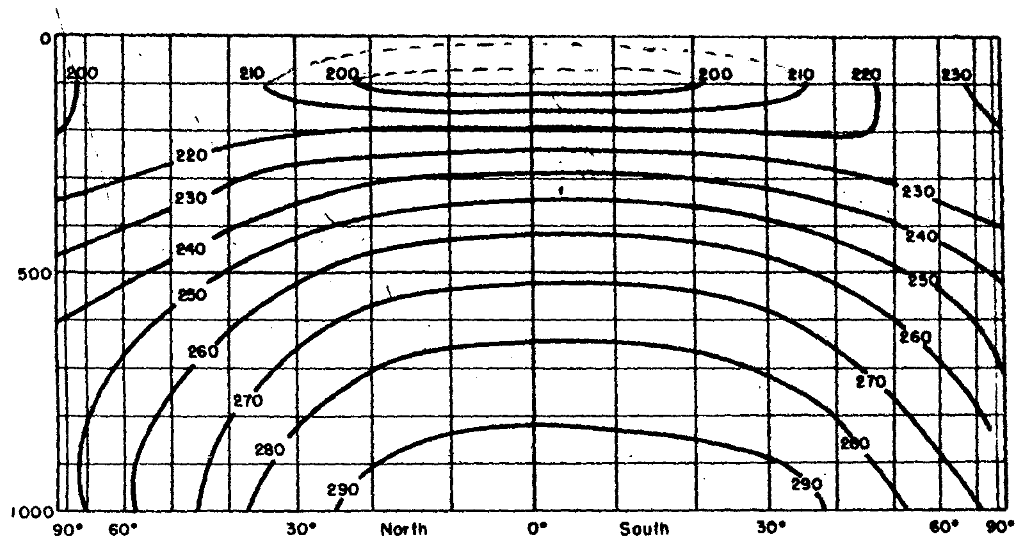


Figure 3.2(a) The time-and-longitude averaged temperature $[\overline{T}]$ in January as estimated by Palmén and Newton (1967). Values are in degrees K

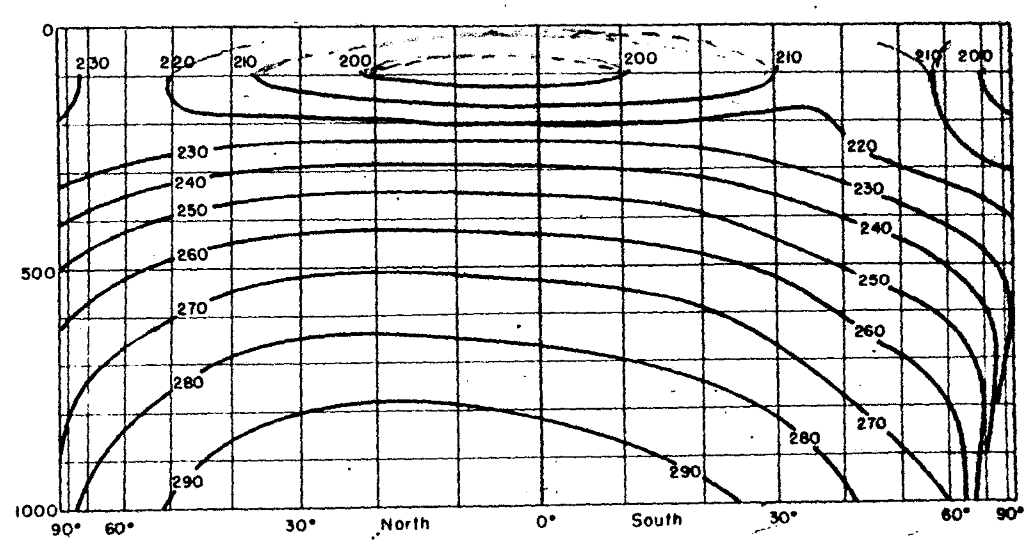


Figure 3.2(b) The time-and-longitude averaged temperature $[\overline{T}]$ in July as estimated by Palmén and Newton (1967). Values are in degrees K

and higher latitudes. The zonally average temperatures are more symmetric with respect to equator during winter than during summer. During winter, highest temperatures in lower and middle troposphere occur near the equator, while during summer, the thermal maximum occurs around 15°-20°N. The meridional temperature gradients at southern higher latitudes during southern winter are also much stronger than those at northern higher latitudes during northern winter.

The vertical variation of temperature has a significant feature of separation of atmosphere into the troposphere, where the temperature decreases with elevation, and the stratosphere, where the temperature no longer decreases and even increases with height. The layer separating these two temperature regimes is called tropopause, and is not sharply defined in average temperature fields as can be observed in day-to-day vertical temperature profiles.

3.3 PHYSICAL CAUSES FOR OBSERVED ZONALLY AVERAGED FIELDS

The above zonally averaged wind and temperature profiles are derived from daily observation and hence must satisfy the zonally averaged relevant equations of atmospheric flow. The sources and sinks of respective fields also play important role in determining these average fields. We shall outline below the causes for observed zonally averaged wind, and temperature fields on dynamic and thermodynamic considerations.

3.3.1 ZONAL WIND.

The zonally averaged wind is maintained by dynamic as well as thermal factors. Due to molecular and boundary layer friction there is continuous dissipation of the kinetic energy of the atmospheric circulation. Therefore, there must be some physical processes by which the momentum or kinetic energy, is generated nearly at the same rate at which it is being destroyed by dissipative forces. One way to look at the maintanance of mean zonal circulation is from the angular momentum considerations.

The absolute angular momentum of a unit mass of air at a latitude θ is given by

$$M = (u + \Omega a \cos\theta) a \cos\theta \qquad 3.1$$

Now there are two forces, i.e., the pressure gradient forces and frictional force which can change the angular momentum. Therefore, the torque of pressure gradient force and frictional force must be equal to the acceleration of angular momentum, or

$$\frac{dM}{dt} = \left(-\frac{1}{\rho}\frac{\partial p}{\partial x} + F_x\right) a\cos\theta \qquad 3.2$$

which, using the continuity equation 2.1(e), may be written as

$$\frac{\partial}{\partial t}(\rho M) = -\nabla.(\rho M V) - \frac{\partial}{\partial x}(\rho a\cos\theta) + \rho a\cos\theta . F_x \qquad 3.3$$

On decomposing the wind components into their zonal mean, and deviation from them, the above equation when integrated over a volume enclosed by vertical walls along latitude circles θ_1 and θ_2 extending from earth's surface to the top of the atmosphere, takes the form

$$\int \frac{\partial}{\partial t} M \rho \, dv = 2\pi a^2 \cos^2\theta \int_{\theta_1}^{\theta_2}\int_0^{\infty} \left(\Omega \bar{u} a\cos\theta + \bar{u}\bar{v} + \overline{u'v'}\right) dz \, d\theta$$

$$- \int \frac{\partial p}{\partial x} a\cos\theta \, dv + \int \rho a\cos\theta F_x \, dv \qquad 3.4$$

The second term also vanishes except when there are large mountains along a latitude. In that case, it is equal to the mountain torque due to difference in pressure on east and west sides of mountain. The last term is frictional dissipation term. The three components of the first term represent respectively the meridional transport of earth's momentum by mean meridional wind ($\Omega \bar{v} a \cos\theta$), the meridional transport of mean westerly momentum by mean meridional flow ($\bar{u}\bar{v}$), and meridional transport of zonal momentum by eddies ($\overline{u'v'}$). These three processes are largely responsible for maintaining the observed mean zonal wind field. Fig. 3.3 shows the relative magnitude of angular momentum fluxes due to these three components between 20° and 30°N as computed by Riehl and Yeh (1956). Of the three components meridional transport of earth's momentum is maximum and is known as the cause of mid-latitude westerlies. The eddy transport of angular momentum has also been found to play a great role in maintaining the westerly flow, particularly the westerly jet. Figure 3.4(a) and (b) shows the meridional transport of westerly momentum by eddies during summer and winter seasons after Oort and Rasmusson

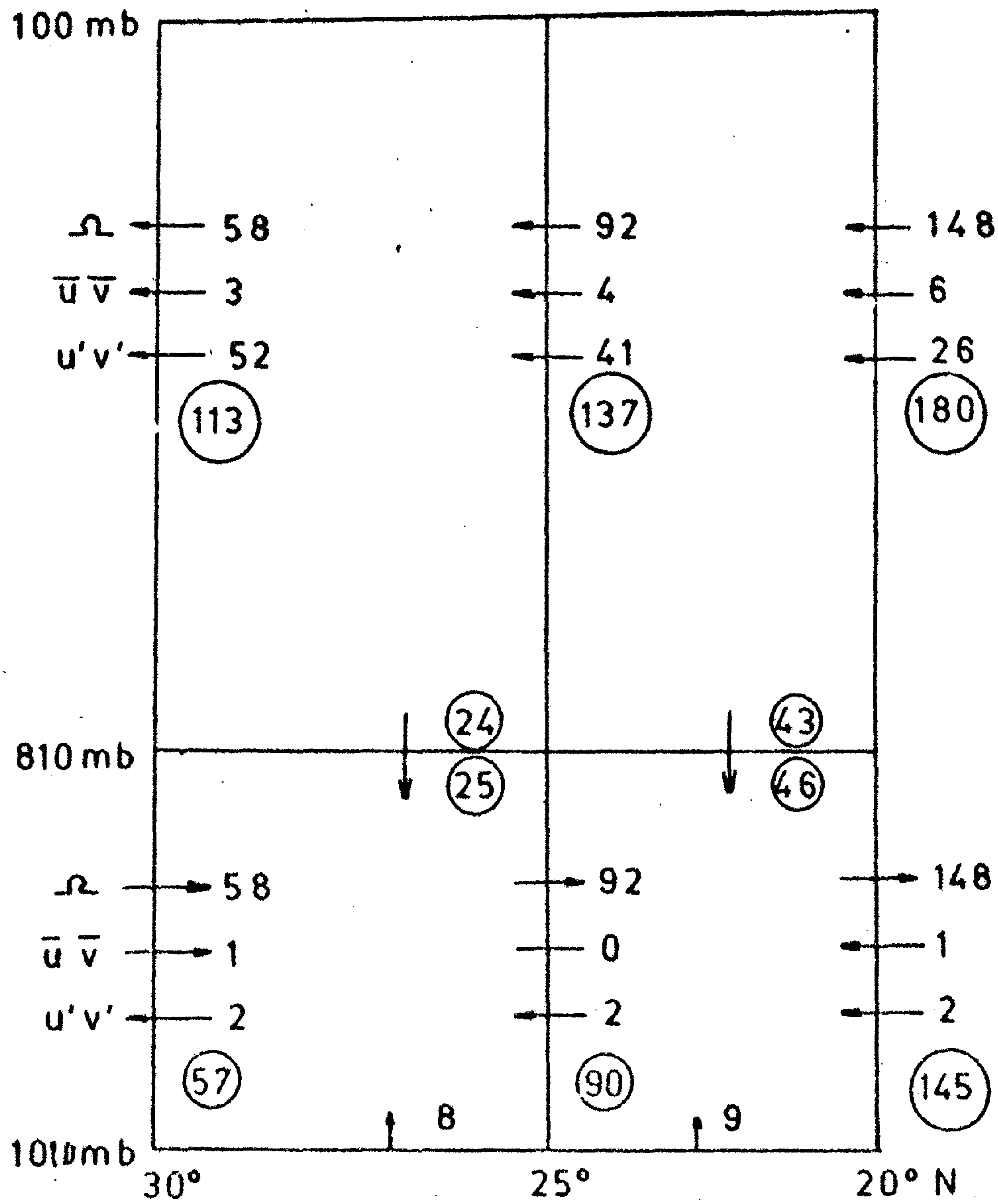


Fig. 3.3 Angular momentum balance for the latitude belt 20°N to 30°N in January (Riehl and Yeh 1954)

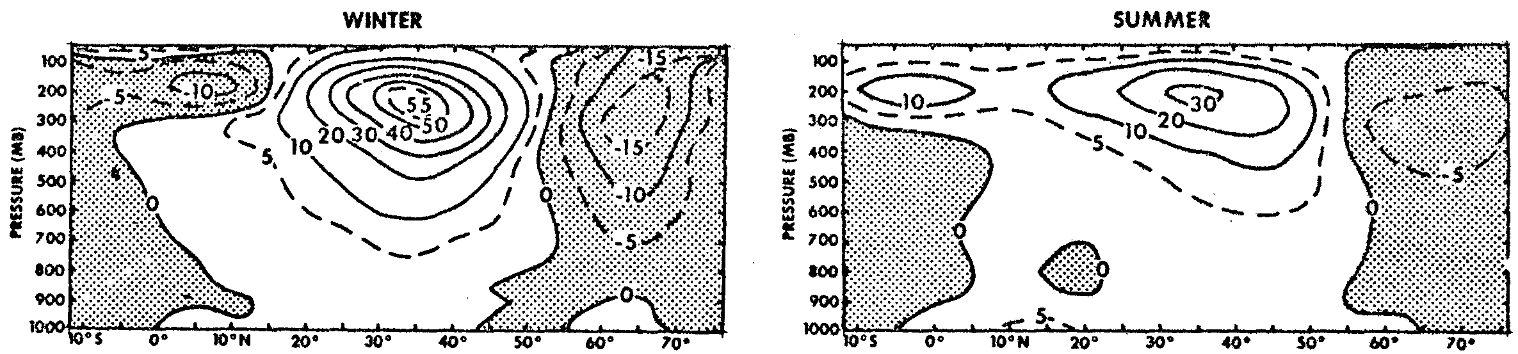


FIGURE 3.4—Northward transport of westerly momentum by all eddies $[\overline{v'u'}]+[\bar{v}^*\bar{u}^*]$ for year and seasons. (Units in $m^2\ sec^{-2}$.)

(1971). We see that during both the seasons the region of maximum meridional transport of westerly momentum by eddies coincides with the location of westerly jet stream (Fig. 3.1(a) and (b)). During winter season the meridional transport of westerly momentum by eddies is nearly twice of that during the winter. The strength of the westerly jet during the two seasons also has similar variations. Thus, the meridional transport of angular momentum plays an important role in maintaining the mid-latitude westerlies and their jet. The meridional transport of relative westerly momentum ($\bar{u}\bar{v}$) is, in the mean, rather small.

Another factor responsible for westerly jet is the sharp equator to pole temperature gradient at middle latitudes. From the thermal wind considerations, the westerlies should increase with height till the tropopause, and decrease thereafter as, on higher levels, the temperature gradient becomes very weak or even reverses.

The angular momentum so generated at a latitude by meridional fluxes and thermal effects is ultimately destroyed by frictional forces nearly at the same rate. This results in maintenance of zonal wind field.

3.3.2 TEMPERATURE FIELD.

The thermodynamic equation may be written as

$$\frac{\partial T}{\partial t} = -\left[u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y}\right] - \frac{\omega}{\theta_T}\frac{\partial \bar{\theta}_T}{\partial p} + \frac{1}{C_p}\dot{Q} \qquad 3.5$$

using the continuity equation and integrating (3.5) from bottom of the atmosphere to its top between ~~low~~ two latitude circles θ_1 and θ_2 we get at any pressure level.

$$\frac{\partial \bar{T}}{\partial t} = -\int_{\theta_1}^{\theta_2}\frac{1}{a}\frac{\partial \overline{Tv}}{\partial \theta}\,d\theta + \overline{\omega\left(\frac{\partial T}{\partial p} + \frac{T}{\bar{\bar{\theta}}_T}\frac{\partial \bar{\theta}_T}{\partial p}\right)} + \overline{\frac{1}{C_p}\dot{Q}} \qquad 3.6$$

where a bar indicates latitudinal mean. `T' and `v' in the first term may be decomposed into zonal mean and eddies so that equation (3.6) takes the form

$$\frac{\partial \bar{T}}{\partial t} = -\int_{\theta_1}^{\theta_2}\frac{1}{a}\frac{\partial \bar{T}\bar{v}}{\partial \theta}\,d\theta - \int_{\theta_1}^{\theta_2}\frac{1}{a}\frac{\partial \overline{T'v'}}{\partial \theta}\,d\theta + \overline{\bar{\omega}\left(\frac{\partial T}{\partial p} + \frac{T}{\bar{\theta}_T}\frac{\partial \bar{\theta}_T}{\partial p}\right) + \frac{1}{C_p}\dot{Q}}\Big)$$

$$\qquad 3.7$$

The terms on the right hand side indicate, respectively, the contribution of

i) mean meridional thermal flux

ii) thermal fluxes by eddies

iii) vertical fluxes of temperature field, and

iv) zonally averaged external sources and sinks of heat.

If we look at the zonally averaged heat budget of the earth's atmosphere (Houghton, 1954) (Fig. 3.5) we see that in the northern hemisphere, tropical and subtropical regions upto 30°N receive more heat by incoming solar radiation than they lose the same by long wave radiation. Over the higher latitudes on the other hand, loss of heat by long wave radiation is more than the gain by incoming solar radiation. Thus, the tropics should have a net surplus heating while higher latitudes should have net cooling. However, over long periods, the temperature at various latitudes remain nearly unchanged. This is accomplished by meridional transport of heat from tropics to mid-latitudes. Out of various transport processes the meridional transport of heat by eddies is largely responsible for removing excess heat for tropics and

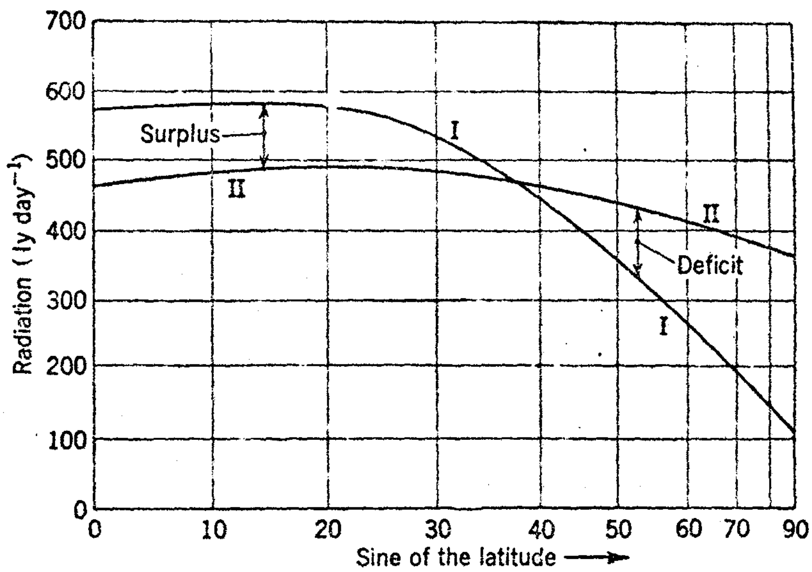


FIG.3·5 Curves I and II represent mean annual insolation and outgoing long-wave flux, respectively, at the tropopause. [*After Houghton* (11).]

transferring it to higher latitudes so as to maintain the observed zonally averaged temperature field. Fig. 3.6(a) and (b) shows the meridional transport of heat by eddies during summer and winter seasons. We see that during winter, heat is transported by eddies towards pole, the strongest transport being in the middle latitudes in the lower troposphere, where prevailing equator to pole temperature gradients are strong. During summer season, there is small southward transport of heat between equator and 20°N also. The poleward transport of heat in the middle and higher latitudes is much smaller during summer than during winter.

The cause for reversal of vertical temperature gradient near the tropopause is both dynamic and photochemical. The ozone in the stratospheric regions absorb ultra-violet radiation and thus warm these regions. Besides, the convective processes in the troposphere change vertical temperature profile in such a way, that we get a reversal of temperature gradient around 200 mb level which is the level of tropopause.

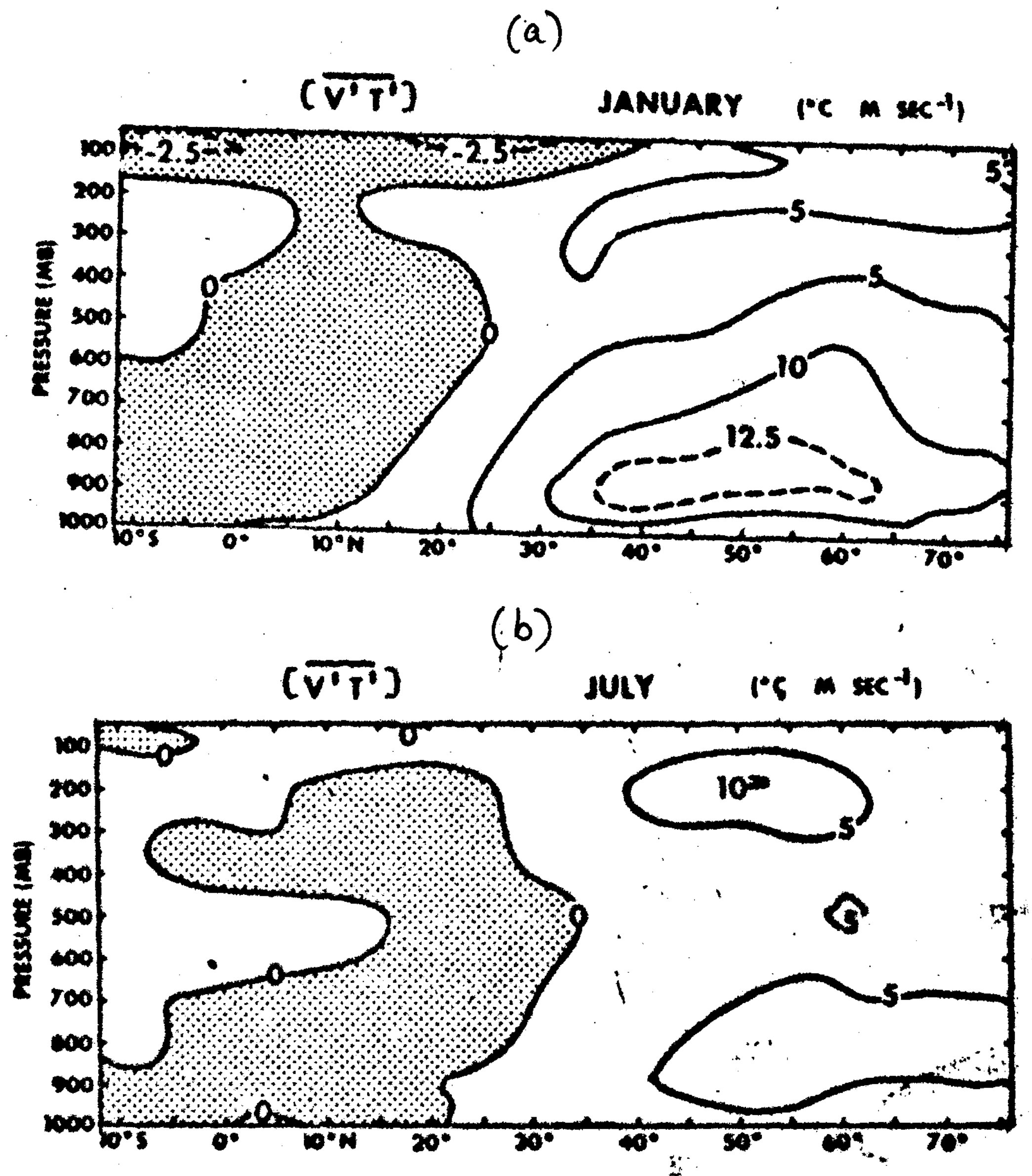


FIG. 3.6 Meridional transport of heat by eddies during summer and winter seasons (Oort, 1964)

3.4 SUMMARY

We have examined above only the zonally averaged features of wind and temperature fields and their causes. We find that both the westerly angular momentum and net heat input in the tropics is more than those in the higher latitudes. Unless some mechanism exists to transport the excess of the westerly momentum and the heat, the observed nearly steady state zonally averaged wind and temperature fields will be impossible to maintain. The meridional transport of angular momentum and heat is accomplished both by the mean meridional circulation and by the eddy fluxes. The westerly jet stream is maintained both by the convergence of angular momentum by eddies and the effect of meridional temperature gradient through thermal wind equations. The eddies also transport heat from the lower latitudes to the higher latitudes so as to maintain heat balance of the earth-atmosphere system. Thus, the earth atmosphere system has developed an inbuilt mechanism by which the momentum and heat are transported from the excess to deficit regions and are simultaneously destroyed by frictional and radiative effects. The rate of generation and dissipation of momentum and heat energy are in near equilibrium, so that in the long run, except for seasonal variations, there is no net gain or loss in the momentum or heat budget of earth-atmosphere system, including of course the oceans which we have not considered in our study.

CHAPTER IV

ENERGETICS OF THE ATMOSPHERE

4.1 INTRODUCTION

An important aspect of studying the atmospheric system is through analysing its energy processes. The atmosphere possesses many forms of energy, like potential energy, kinetic energy, electrical energy and photo-chemical energy etc. The energies like electric energy or photo-chemical energy are of much smaller magnitude and also exist over limited domain of the atmosphere, or over limited periods. In the general circulation studies, our concern with the atmosphere energetics relate to three forms of energy only, i.e.,

i) Potential energy (P) due to the position of the air particle in relation to some reference level,

ii) The internal energy (I) due to the temperature of the air parcel, and

iii) The kinetic energy (K) due to the motion of the atmosphere.

The potential energy is given by the expression

$$P = gz$$

and the internal energy by

$$I = C_v T$$

It can be shown that u

the P and I of entire

earth's surface to the

constant ratio or more

$$P = \frac{R}{c_v} I$$

From (4.1) we see that

$$P + I = \frac{c_p}{R} P =$$

Thus, under hydrostati

column of air are not

relations (4.1) or (4.

to regard P and I as a

Margules (1903) as the

With this considerati

of energy, i.e.,

i) The Total Poten

ii) The Kinetic Ene

to form the subject th

Both the observ

considerations have sh

energy is much larger

of the atmosphere, their relative magnitude being the order of 2000: 1. Thus, the study of quantities having so largely different magnitudes could not be free from errors of relative approximation. Actually there is a large amount of total potential energy which does not play any role in the kinetic energy balance of the atmosphere. Based on these considerations, Lorenz (1955) put forth the concept of Available Potential Energy (A).

4.2 AVAILABLE POTENTIAL ENERGY

According to Lorenz (1955), available potential energy is that amount of potential energy which at any time is available for conversion to kinetic energy. The available potential energy of the atmosphere is completely determined by the distribution of its mass. If the mass distribution is such that the atmosphere has a statistically stable horizontal stratification, which is possible if the pressure surfaces coincide with the potential temperature surfaces, the potential energy of the atmosphere is minimum. Thus, the available potential energy is the difference between the actual potential energy of the atmosphere and potential energy which the atmosphere will acquire if its pressure surfaces and temperature surfaces are brought parallel to each other. Based on these considerations, Lorenz (1955) derived the expression for available potential energy (A) as

$$\bar{A} = \frac{1}{2}\int_0^{\infty} \bar{T}\left(\Gamma_d - \Gamma\right)^{-1}\left(\frac{T'}{\bar{T}}\right)^2 dp \qquad 4.3(a)$$

or $$\bar{A} = \frac{1}{2}\,\frac{R}{g}\,p_{00}^{-K}\int_0^{\infty} \bar{\theta}_T^{-2}\, p^{-(1-K)}\left(-\frac{\partial\bar{\theta}_T}{\partial p}\right)\left(\frac{\theta_T'}{\bar{\theta}_T}\right)^2 dp \qquad 4.3(b)$$

The expression (4.3) shows that the available potential energy is given by the variance of temperature or the potential temperature, on a constant pressure surface. If the temperature on a pressure surface is constant, the variance of thermal field is zero and hence the atmosphere at that level possesses no available potential energy. The scale analysis considerations show that the ratio of mean available potential to mean total potential energy is

$$\bar{A}/(\bar{P}+\bar{I}) \sim 1/2000$$

and the ratio between kinetic energy and available potential energy is

$$\bar{K}/\bar{A} \sim 1/10$$

Based on these magnitude considerations the studies of atmospheric energetics generally relate to the available potential energy and kinetic energy, and their conversion and exchange processes.

4.3 EQUATIONS FOR STUDY OF ATMOSPHERIC ENERGETICS

The basic equations for studying the atmospheric energetics are the set of equations (2.1) or (2.2) given in Chapter II. However, we manipulate these equations in such a form that they become the equations for the kinetic energy and the available potential energy of the atmosphere. The horizontal momentum equations of the atmosphere in pressure co-ordinate system as given in Chapter II are:

$$\frac{\partial u}{\partial t} = -\left[u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + \omega\frac{\partial u}{\partial p}\right] - fv - \frac{\partial \phi}{\partial x} + F_x \qquad 4.4(a)$$

$$\frac{\partial v}{\partial t} = -\left[u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + \omega\frac{\partial v}{\partial p}\right] + fu - \frac{\partial \phi}{\partial y} + F_y \qquad 4.4(b)$$

Multiplying 4.4(a) by 'u' and 4.4(b) by 'v' and adding the resulting two equations, we get the equation for the kinetic energy integrated over a specified domain as:

$$\frac{\partial K}{\partial t} = -\frac{1}{g}\int_0^{p}\overline{\Big[\underbrace{\nabla . K\vec{V}}_{(A)} + \underbrace{\nabla\phi . \vec{V}}_{(B)}\Big]}dp - \frac{1}{g}\int_0^{p}\underbrace{\overline{\omega\alpha}}_{(C)}\, dp + \frac{1}{g}\int_0^{p_0}\underbrace{(uF_x + vF_y)}_{(D)}\, dp \qquad 4.5$$

The thermodynamic equation is given by

$$\frac{\partial \theta_T}{\partial t} = -\Big[u\frac{\partial \theta_T}{\partial x} + v\frac{\partial \theta_T}{\partial y} + \omega\frac{\partial \theta_T}{\partial p}\Big] + \frac{1}{C_p}\frac{\theta_T}{T}\dot{Q} \qquad 4.6$$

Breaking θ_T and $\dot{Q}$ into its zonal averages and deviations, and presuming that the zonally averaged θ_T and $\dot{Q}$ do not change with time, the above equation may be written as:

$$\frac{\partial \theta_T'}{\partial t} = -\Big[u\frac{\partial \theta_T'}{\partial x} + v\frac{\partial \theta_T'}{\partial y}\Big] - \omega\frac{\partial \bar{\theta}_T}{\partial p} + \frac{1}{C_p}\Big(\frac{P_{00}}{p}\Big)^{k}\dot{Q}' \qquad 4.7$$

Multiplying both sides of (4.6) by θ_T' and integrating the equation over a specified horizontal domain from the bottom to the top of the atmosphere, we get after some manipulations, the equation for available potential energy as:

$$\frac{\partial \bar{A}}{\partial t} = \int_0^{A}\underbrace{\overline{(\vec{V} . \nabla A)}}_{(E)}dp + \frac{1}{g}\int_0^{p_0}\underbrace{\overline{\omega\alpha}}_{(F)}\, dp + \frac{C_p}{g}\int_0^{p}\underbrace{\theta_T'\dot{Q}'}_{(G)}\, dp \qquad 4.8$$

In equation (4.5), terms (A) and (B) represent horizontal fluxes of kinetic energy and potential energy. The term (D) represents the dissipation of kinetic energy.

In equation (4.8) the term (E) represent~~s~~ the advection of available potential energy and the term (G) its generation or dissipation.

Term (C) in equation (4.5) for kinetic energy and term (F) in equation (4.8) for available potential energy, are equal but have opposite signs. This term, therefore, represents the conversion of available potential energy to kinetic energy.

We shall look into these terms in a more detailed way. We see that the term determining the conversion of available potential energy to kinetic energy is given by $\int_0^p \overline{\omega\alpha}\, dp$ or it is the variance between vertical velocity and the specific volume. If the air rises over warm regions or sinks over cold regions, this term will be negative and will indicate decreasing available potential energy, and a corresponding increase in kinetic energy. Thus, the conversion of available potential energy to kinetic energy takes place by rising of warm air and/or sinking of cold air. Such processes take place in direct Hadley circulations, where warm air rises over tropics and cold air sinks over the mid and higher latitudes. If the cold air rises and warm air sinks, the energy conversion is in an opposite direction, i.e., from kinetic energy to available potential energy.

The term (G) in (4.8) representing the generation or dissipation of applied potential energy is a measure of co-variance between potential temperature and diabatic heating. Thus, if the warm regions get more heated or the cold regions become more cool, this term will be positive and there will be a generation of available potential energy. The warming of cold regions and the cooling of warm regions, on the other hand, leads to decrease or dissipation of available potential energy.

Term (D) in equation (4.5) representing the dissipation of kinetic energy, generally contributes towards decrease in kinetic energy. It is possible to further split the eddy motion into various scales represented by Fourier Components (Saltzman 1957; Saltzman and Fleisher 1960; Murakami and Tomatsu 1965). Such break up can provide insight into the energy exchange processes and energy conversion processes between eddies of different scales and the zonal flow, as well as among eddies themselves.

4.4 OBSERVED FEATURES OF KINETIC ENERGY AND POTENTIAL ENERGY DISTRIBUTIONS

Table (4.1) gives the values of Total Available Potential Energy (A), Zonally Averaged Available Potential Energy (AZ), the Eddy Available Potential Energy (AE), Total Kinetic Energy (K) and Zonal Kinetic Energy (KZ) and Eddy Kinetic Energy (KE) for four representative months January, April, July and October 1962 after Wiin-Nielsen (1965).

Table (4.2) gives the ratios AE/AZ, KE/KZ, KE/AE, KZ/AZ K/A for the values in Table (4.1). Although various energy statistics may differ with different periods of data to some extent, the above statistics may be regarded as representative for the northern middle latitudes.

We see that the kinetic energy (K) is about half of the available potential energy (A) during all the four months. The K to A ratio in this case is somewhat higher than the long term expected ratio of 0.1. We notice that the eddy available potential energy (AE) is from 25% to 40% of zonal available potential energy (AZ). Eddy kinetic energy (KE) on the other hand is larger than the zonal kinetic energy (KZ). It is also interesting to note that the zonal kinetic energy (KZ) is a small

Table 4.1: Monthly Mean Available Potential Energy values during 1962 (units Kjm^{-2})

Month	A	AZ	AE	K	KZ	KE
January 1962	5376	4054	1322	3013	1340	1673
April 1962	4006	3202	804	2014	798	1216
July 1962	1952	1356	596	1002	317	685
October 1962	3916	3046	870	1810	662	1148
Mean	3813	2915	898	1959	779	1180

Table 4.2: Energy Ratios

Month	AE/AZ	KE/KZ	KZ/AZ	KE/AE	K/A
January 1962	0.33	1.25	0.33	1.27	0.56
April 1962	0.25	1.52	0.25	1.51	0.50
July 1962	0.44	2.16	0.23	1.15	0.51
October 1962	0.29	1.73	0.22	1.32	0.46
Mean	0.31	1.51	0.27	1.31	0.57

fraction of zonal available potential energy (AZ) while eddy kinetic energy (KE) is larger than eddy available potential energy (AE).

From above, we see that whereas the available potential energy is largely concentrated in its zonal component, the kinetic energy has larger magnitudes in the eddy component of the flow.

When we look into the kinetic energy and available potential energy distribution into wave number domain we get further details. Figure 4.1 gives the distribution of kinetic energy over wave number 1 to 15 during summer and winter seasons of 1969. We see that the kinetic energy during both the seasons decreases with wave number. However, ? during winter there appears to be a weak maximum at wave number 4.

Figure 4.2 shows the ratio of spectral distribution of available potential energy during summer and winter seasons. In this case, we see that for large scale waves upto wave number 7, the summer season available potential energy is more than that during the winter. For higher wave numbers, the available potential energy during winter is larger than during summer.

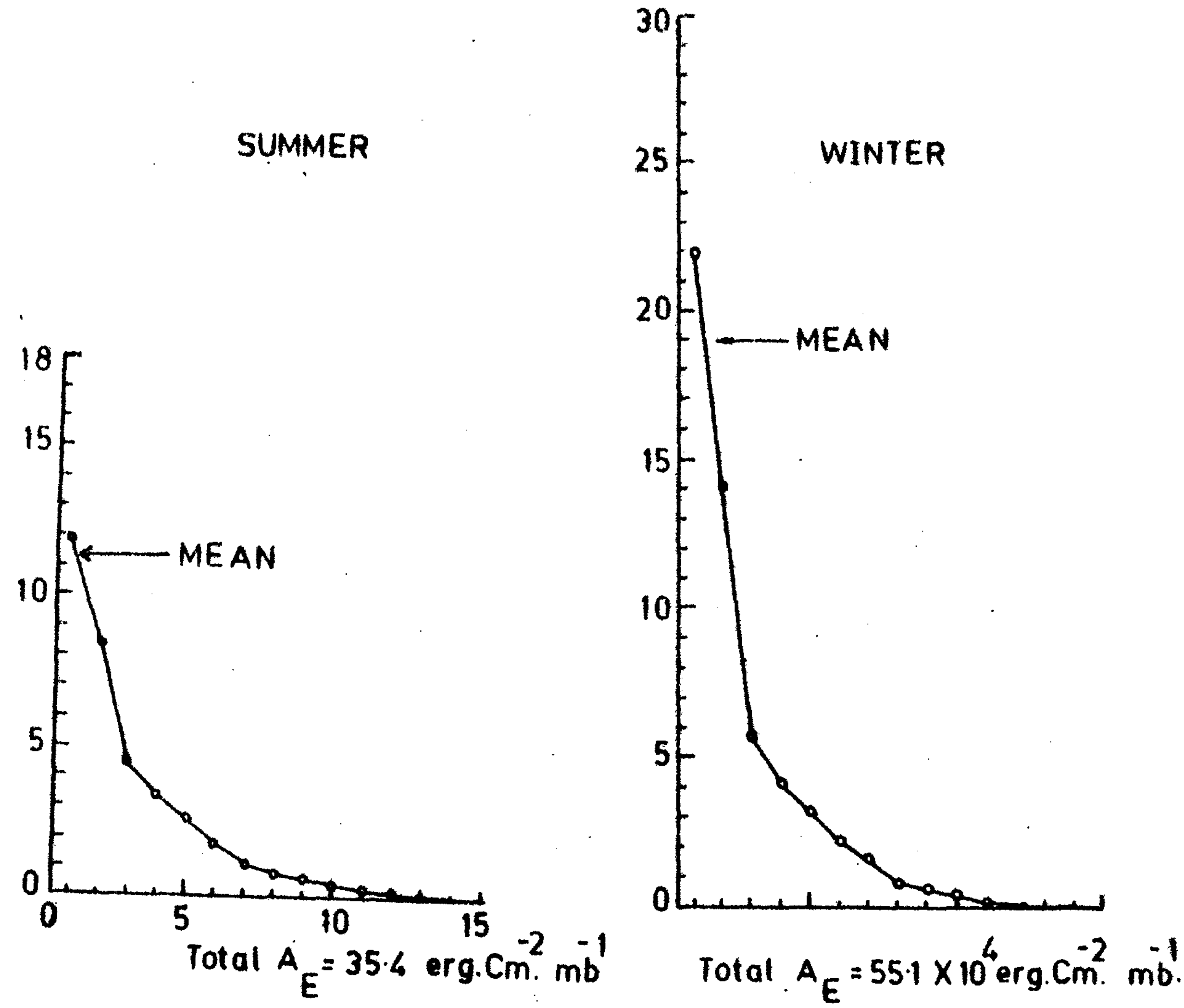


FIG. 4.1 Distribution of kinetic energy during Summer and winter of 1969 (Bedi, 1976)

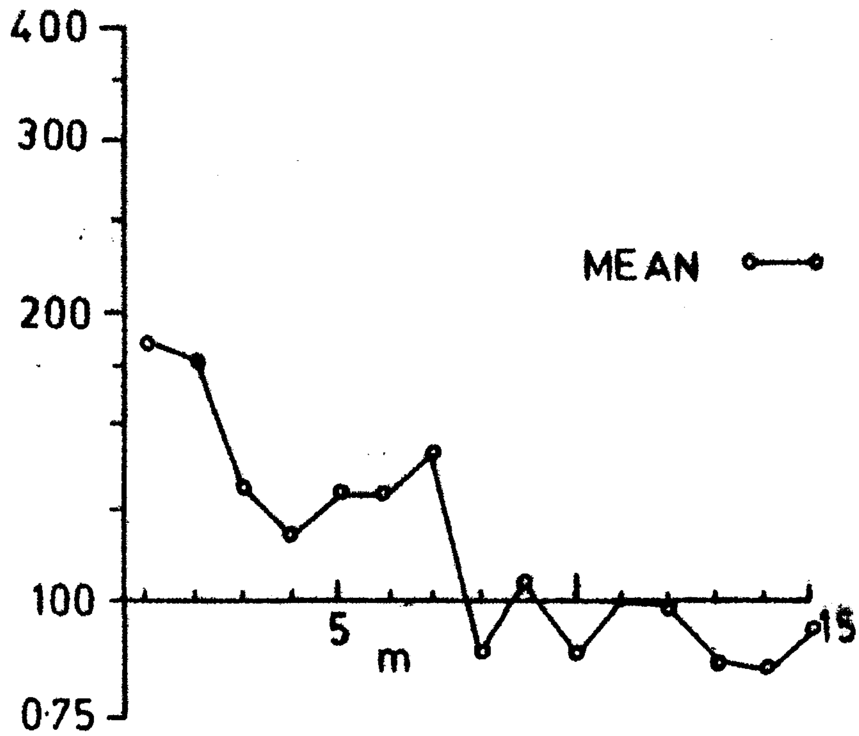


FIG. 4.2 Ratio of distribution of available potential energy during summer and winter of 1969 (Bedi, 1976)

4.5 ENERGY EXCHANGE AND CONVERSION PROCESSES

A number of studies have been performed to study the energy conversion and exchange processes in the past (Murakami and Tomatsu 1966; Saltzman and Fleisher 1962; Wiin-Nielsen et al. 1964, 1967; Oort 1964). All these studies have, by now, established some well defined features of atmospheric energetics. For the troposphere the results of these studies can be summarized as in Figure 4.3 based on Oort (1964a). According to this, the atmospheric energy process in the troposphere is as follows :

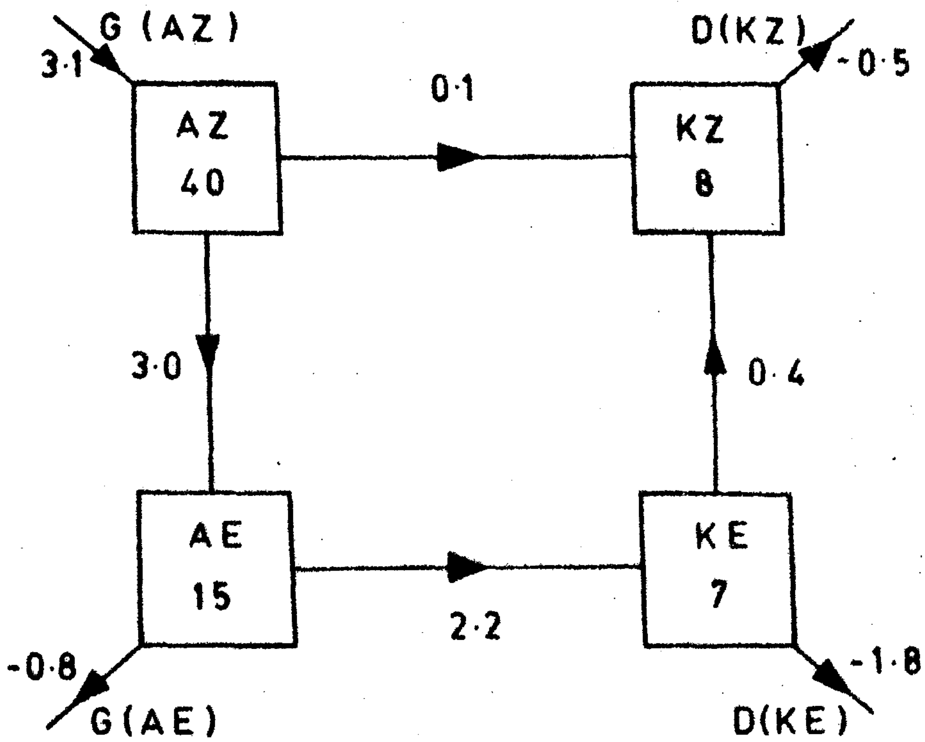
i) Due to solar insolation there is a generation of zonal available potential energy.

ii) A larger part of zonal available potential energy is converted into eddy available potential energy which, in turn, is converted into eddy kinetic energy.

iii) The eddy kinetic energy is then converted into the zonal kinetic energy.

iv) The direct conversion of zonal available potential energy to zonal kinetic energy is rather very small.

v) The dissipation of kinetic energy as well as of available potential energy is largely of their



Energy units are in 10 Jm^{-2} and energy transformation units are in Watt m^{-2}

Fig. 4·3. Atmospheric Energy cycle in the troposphere (Oort. 1964a) eddy components. The kinetic energy dissipation also occurs in the zonal field.

The energy cycle in the stratosphere is very different and is shown in Figure 4.4 (after Oort 1964b). The stratospheric energy cycle may be described as follows:

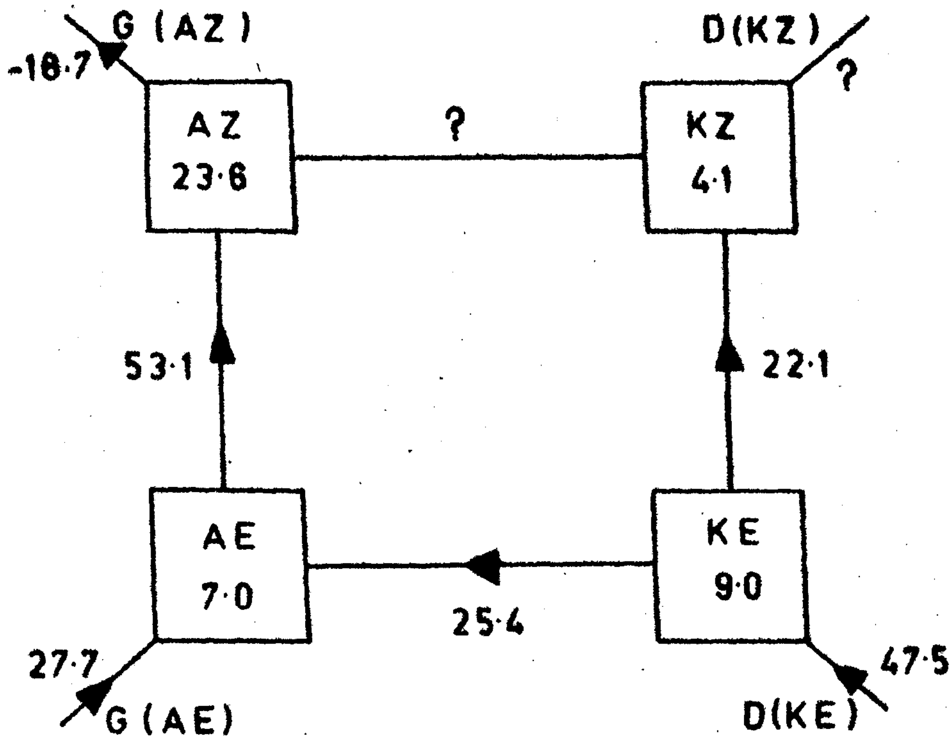
i) The stratosphere receives eddy kinetic energy as well as eddy available potential energy.

ii) The eddy kinetic energy is converted into eddy potential energy, which then is converted into zonal potential energy.

iii) A part of eddy kinetic energy is converted into zonal kinetic energy also.

iv) The direction of conversion between zonal available potential energy and zonal kinetic energy is not clear. The dissipation of available potential energy largely takes place at the zonal available energy.

Thus, the energy cycle in the lower stratosphere is in an opposite direction to that in the troposphere.



Energy units are in 10^{25} erg and energy transformation units are in 10^{18} erg sec^{-1}

Fig. 44. Atmospheric Energy cycle in the Stratosphere - northern hemisphere 100 - 30 mb layer (Oort. 1964b)

4.6 SUMMARY

From the above discussion, we notice the following features of atmospheric energy processes:

i) The available potential energy is much larger in magnitude than the kinetic energy of the atmosphere. However, the eddy kinetic energy is larger in amount than the eddy available potential energy. Thus, the available potential energy largely resides in zonally averaged field, whereas the kinetic energy is largely in the eddy motions.

ii) The energy cycle in the troposphere is such that the troposphere behaves like a heat engine. The available potential energy is generated in its zonal component due to heating by sun. This is then converted into eddy available potential energy, which in turn, converts into eddy kinetic energy and finally to zonal kinetic energy. Dissipation of kinetic energy takes place both at its eddy and zonal components, while dissipation of available potential energy takes place from its eddy field.

iii) The stratospheric energy cycle is such that the atmosphere works as a thermodynamic refrigerator.

The source of energy for this region is the eddy kinetic energy and eddy available potential energy, which have been found to be transported from the troposphere. The eddy kinetic energy gets converted into eddy available potential energy which in turn is converted into zonal available potential energy, which ultimately is dissipated due to radiative processes in the stratosphere.

CHAPTER V

SPECTRAL DISTRIBUTION OF THE KINETIC ENERGY IN NORTHERN TROPICS AND SUBTROPICS DURING ACTIVE AND WEAK MONSOON SPELLS

5.1 INTRODUCTION

The wave like nature of global distribution of wind and temperature fields which varies both in space and time, implicates the presence of different scales of atmospheric motion and their role in maintaining the atmospheric circulation. These wave like patterns suggest that the flow may be resolved into a spectrum of spatial scales represented by harmonic components. Expressions can then be derived for calculating kinetic energy and available potential energy of these harmonic components. Saltzman (1957) derived the formulae for energy exchange processes in Fourier domain which involved both the available potential energy (A) and kinetic energy (K). Based on this formulation, a number of studies regarding the energy distribution and the energy exchanges among different scales of motion have been made (Saltzman and Fleisher, 1962; Wiin-Nielsen, 1950; Yang, 1967). An exhaustive survey o. such studies has been given by

Salzman (1970). A number of these studies have used sufficiently large samples of data and provide reliable statistics of energy exchanges upto wave number 15.

The primary conclusions drawn from these studies are, that nearly all the components of eddy motion transfer kinetic energy to the zonal motion. Also the wave number 5-10 (cyclone-scale waves) and the wave number 2 (large scale wave) act as sources of kinetic energy for the other wave components. The eddy available potential energy is converted into eddy kinetic energy nearly at all scales of motion through baroclinic processes, though wave numbers 5-10 appear to derive maximum kinetic energy through this process and transfer it both to the lower and higher wave numbers. As far as potential energy is concerned, the zonal available potential energy is the source of eddy available potential energy for all scales of motion.

5.2 PURPOSE OF THE PRESENT STUDY:

The studies made so far on the spectral distribution of atmospheric energy and its exchange processes have been aimed at deriving climatological features of these processes and their seasonal variation for the middle and higher latitudes. A few of the studies made for low latitudes relate to upper troposphere (Krishnamurti, 1978). The atmospheric wave due to the momentum and heat transport processes associated with them play an important role in influencing the large scale tropical circulation like monsoons. The monsoon season is not a continuous spell of good rain, but there are periods of heavy rain, poor rains or drought conditions and normal rainfall activity within the same monsoon season. There could be many factors, both of local and global extent, which are responsible for such variations in the monsoon. It is now well known that the monsoon circulation is associated with zonal wave number 2, as a result of differential heating associated with two continental (Eurasia and America) and two ocean (the Atlantic and the Pacific) systems. The purpose of the present study is to look into the spectral distribution of mid-tropospheric kinetic energy during a spell of active monsoon and a spell of weak monsoon. The monsoon activity is also influenced by the circulation in mid-latitudes. The latitudinal belt selected for this study therefore, extends from 5°N to 45°N.

5.3 FOURIER ANALYSIS OF GEOPOTENTIAL HEIGHT FIELD

Before calculating the distribution of kinetic energy into various wave components, the geopotential height field at each latitude was decomposed into Fourier components upto wave number 15. The Fourier analysis is based on the following orthogonality properties of sine and cosine functions:

$$\int_0^{2\pi} \sin m_1\lambda \sin m_2\lambda \, d\lambda = \pi \quad , \quad m_1 = m_2 \qquad 5.1(a)$$
$$= 0 \quad , \quad m_1 \neq m_2$$

$$\int_0^{2\pi} \cos m_1\lambda \cos m_2\lambda \, d\lambda = \pi \quad , \quad m_1 = m_2 \qquad 5.1(b)$$
$$= 0 \quad , \quad m_1 \neq m_2$$

$$\int_0^{2\pi} \sin m_1\lambda \cos m_2\lambda \, d\lambda = 0 \quad , \quad m_1 = m_2 \qquad 5.1(c)$$
$$m_1 \neq m_2$$

The Fourier analysis involves the representation of any space variable $h(\lambda)$ in terms of a truncated series of sine and cosine functions as

$$h(\lambda) = a_0 + a_1\cos\lambda + a_2\cos 2\lambda + \ldots\ldots a_n\cos n\lambda$$
$$+ b_1\sin\lambda + b_2\sin 2\lambda + \ldots\ldots\ldots b_n\sin n\lambda \qquad 5.2$$

where the series in equation (5.2) is truncated at wave number n . In our case $n = 15$.

In our case, we have to decompose geopotential height available at 36 grid points along a latitude circle, into Fourier components, so that an equivalent form of relation 5.2 in terms of points 1 to 36 may be written as

$$h_i = a_0 + a_1 \cos\frac{\pi}{18} i + a_2 \cos\frac{2\pi}{18} i + \ldots\ldots a_n \cos\frac{n\pi}{18} i$$
$$+ b_1 \sin\frac{\pi}{18} i + b_2 \sin\frac{2\pi}{18} i + \ldots\ldots b_n \sin\frac{n\pi}{18} i \qquad 5.3$$

The Fourier analysis involves the determination of co-efficients $a_0, a_1, a_2 \ldots a_n$ and $b_1, b_2 \ldots b_n$.

To determine a_0 , both sides of 5.3 are summed over i = 1 to 36, so that

$$a_0 = \frac{1}{36} \sum_{i=1}^{36} h_i \qquad 5.4$$

a_n and b_n are determined by multiplying both sides of 5.3 by $\cos\frac{n\pi}{18} i$ and $\sin\frac{n\pi}{18} i$ and summing up both sides over i = 1 to 36 so that using orthogonality property 5.1, we get

$$a_n = \frac{1}{18} \sum_{i=1}^{36} h_i \cos\frac{n\pi}{18} i \qquad 5.5(a)$$

$$b_n = \frac{1}{18}\sum_{i=1}^{36} h_i \sin\frac{n\pi}{18} i \qquad \text{where } n=1,2,\ldots 15 \tag{5.5(b)}$$

By this process we can determine the cosine and sine co-efficients ($a_1, a_2 \ldots a_n$; $b_1, b_2 \ldots b_n$) of Fourier series.

5.4 KINETIC ENERGY CALCULATIONS IN WAVE NUMBER DOMAIN

The kinetic energy of a unit mass of the atmosphere is given by

$$K = \frac{1}{2}\left(u^2+v^2\right) \tag{5.6}$$

where u and V are the zonal and meridional components of wind.

For calculating u and V from the geopotential height we make use of geostrophic relation (2.4), so that:

$$u = -\frac{1}{fa}\frac{\partial\phi}{\partial\theta} = -\frac{1}{fa}\cdot\frac{\partial}{\partial\theta}\left(a_0+\sum_{n=1}^{15}[a_n\cos n\lambda + b_n\sin n\lambda]\right)$$
$$= -\frac{1}{fa}\left(\frac{\partial a_0}{\partial\theta}+\sum_{n=1}^{15}\left[\frac{\partial a_n}{\partial\theta}\cos n\lambda + \frac{\partial b_n}{\partial\theta}\sin n\lambda\right]\right) \tag{5.7(a)}$$

and

$$V = \frac{1}{fa\cos\theta}\cdot\frac{\partial\phi}{\partial\lambda} = \frac{1}{fa\cos\theta}\cdot\frac{\partial}{\partial\lambda}\left[a_0+\sum_{n=1}^{15}\left(a_n\cos n\lambda + b_n\sin n\lambda\right)\right] \tag{5.7(b)}$$
$$= \frac{1}{fa\cos\theta}\cdot\left[\sum_{n=1}^{15}\left(nb_n\cos n\lambda - na_n\sin n\lambda\right)\right]$$

Thus, the sine and cosine components of zonal components of velocity (u) are given by

$$u_s(n) = -\frac{1}{fa}\frac{\partial a_n}{\partial \theta} \text{ and } u_c(n) = -\frac{1}{fa}\frac{\partial b_n}{\partial \theta} \qquad 5.8$$

and those for the meridional velocity (v) are

$$V_s(n) = -\frac{n a_n}{fa\cos\theta} \text{ and } V_c(n) = \frac{n b_n}{fa\cos\theta} \qquad 5.9$$

It may be noted that the zonally averaged zonal velocity is given by

$$-\frac{1}{fa}\frac{\partial a_0}{\partial \theta}$$

while, zonally averaged meridional velocity is zero.

The kinetic energy averaged over a latitude circle is given by

$$\overline{k_n} = \frac{1}{2}\left[u_s^2(n) + u_c^2(n) + V_s^2(n) + V_c^2(n)\right] \qquad 5.10$$

The actual computations involved the following steps :

i) The given geopotential data along a latitude circle (36 values) was analysed into Fourier components up to wave number 15

ii) The sine and cosine amplitudes of zonal and meridional velocity were calculated for each wave number by relation (5.8) and (5.9),

iii) The spectral distribution of kinetic energy was calculated from the sine and cosine components obtained in (ii) above by relation 5.10.

5.5 DATA AND ANALYSIS FOR KINETIC ENERGY DISTRIBUTION STUDY

An international Monsoon Experiment (MONEX-79) was conducted as regional sub-programme of the First GARP Global Weather Experiment during 1979 monsoon season. The data coverage during this period have been very good. We have, therefore, selected this period for our study.

Two phases of contrasting monsoon activity were selected for our study; an active monsoon phase from August 4 to 10, 1979 and a weak phase during the period August 21 to 27, 1979. Both phases covered a period of one week. The active monsoon week from 4 to 10 August 1979 had the following synoptic features :

> The monsoon trough was active from the beginning of the week. On August 6th a depression formed in the Bay of Bengal which developed into a cyclonic storm on

> 7 August 1979. The system moved in land on 10 August 1979 and weakened. In association with this system the monsoon remained active throughout the week and there was good rain in the country.

The weak monsoon week, 21 to 27 August 1979 had the following features:

> The monsoon trough was very close to the Himalayan foot hills. A relatively strong ridge of high pressure developed over the west coast of India and neighbourhood. No depression formed in the Bay of Bengal during the period. Overall break monsoon conditions prevailed over the country and rainfall was scanty.

The data studied consist of the daily geopotential height field at 500 mb between 5°N and 45°N as available on the northern hemispheric charts prepared by Hydrometeorological Service of the U.S.S.R. The data were manually interpolated from these charts at 5° latitude interval and 10° longitude interval. The data were thoroughly checked before subjecting them to further analysis.

The geopotential heights along each latitude circle were decomposed into Fourier components up to wave number 15. The spectra of zonal and meridional velocity components were then calculated by using geostrophic assumption. From these, the kinetic energy for each component was calculated and mean values for the two periods determined.

5.6 DISCUSSION OF RESULTS

Figures 5.1(a) to (d) show the daily variation of the amplitudes of wave numbers 1, 2, 4, 6 and 8 at the latitudes 10°N, 20°N, 30°N and 40°N. The amplitudes of higher wave numbers were much smaller and therefore are not presented. We see the amplitudes of wave number 1 and 2 are generally larger during the active monsoon period than during the weak monsoon period. Along latitude 20°N, 30°N and 40°N the amplitude of wave number 4 are also larger during active phase than during weak phase. There does not seem to be much difference in the magnitudes of the amplitudes of wave number 6 and 8 during the two phases. The wave number 1 to 4 constitute planetary scale monsoon. Thus the activity of the monsoon on a planetary scale is reflected in the higher magnitudes of these components during the active monsoon phase as compared to the weak monsoon phase.

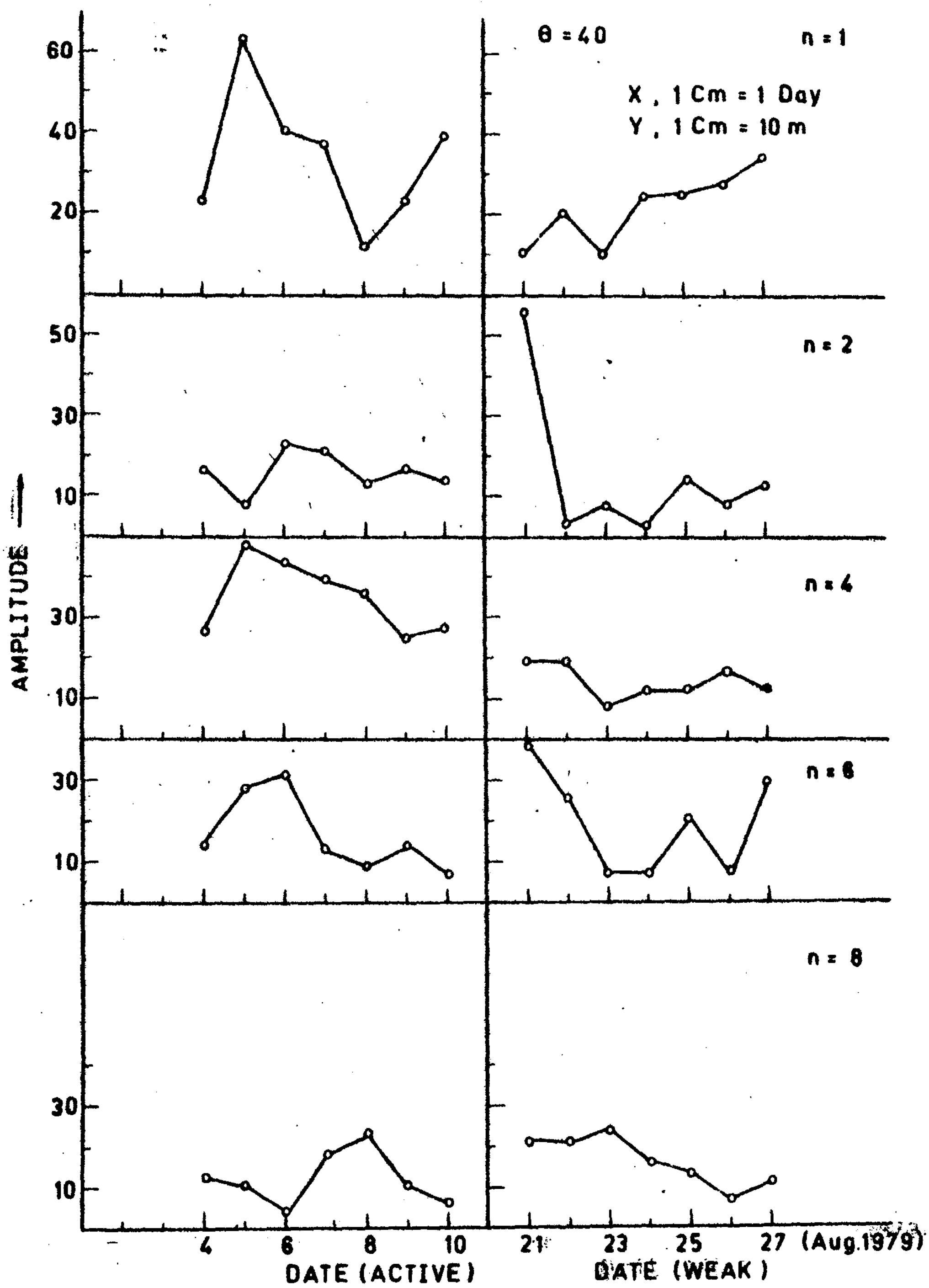


Fig. 5.1(a). Variation of amplitudes of geopotential height field for wave number 1,2,4,6 and 8 during strong and weak monsoon phases.

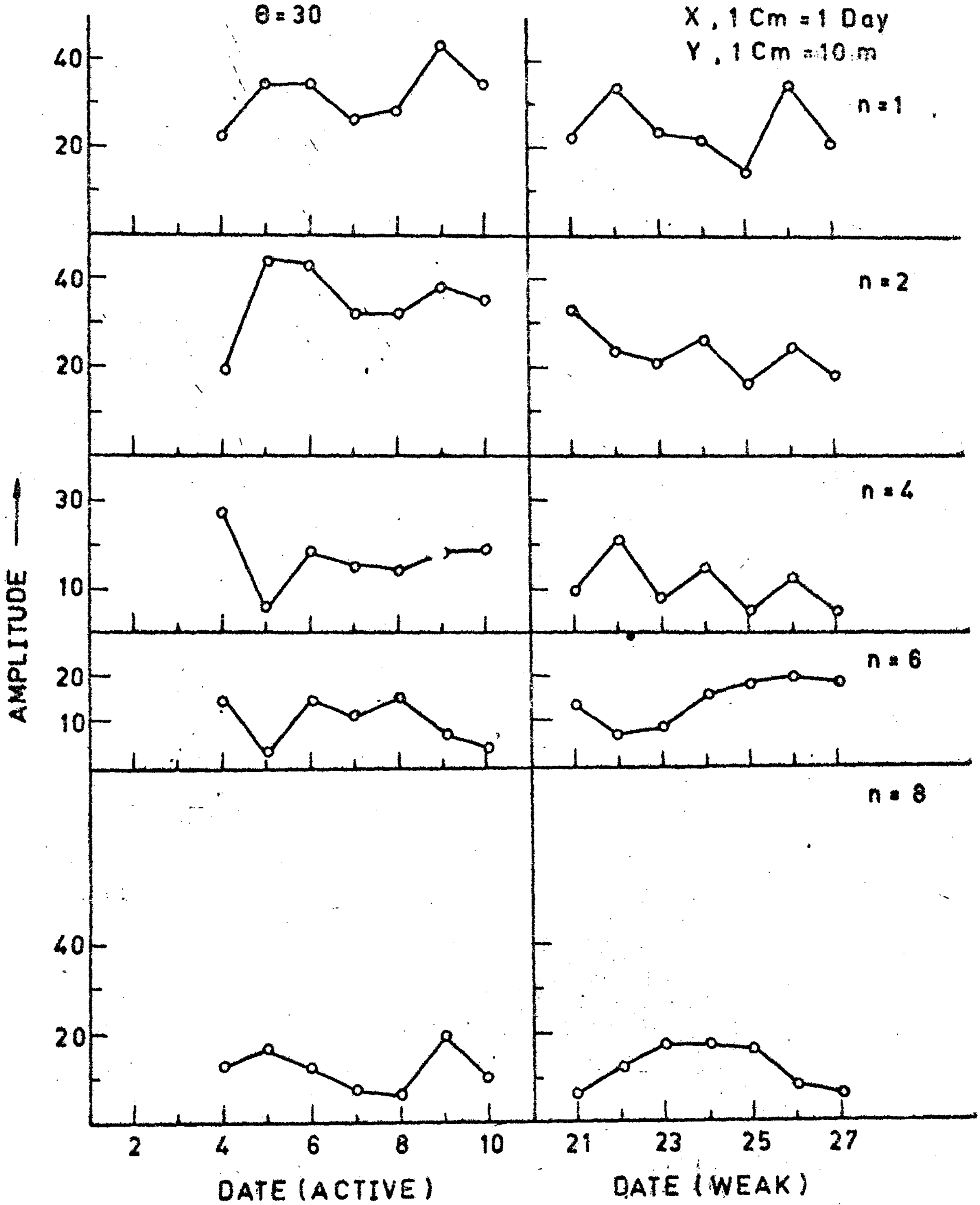


Fig. 5.1 (b). Variation of amplitudes of geopotential height field for wave number 1, 2, 4, 6 and 8 during strong and weak monsoon phases.

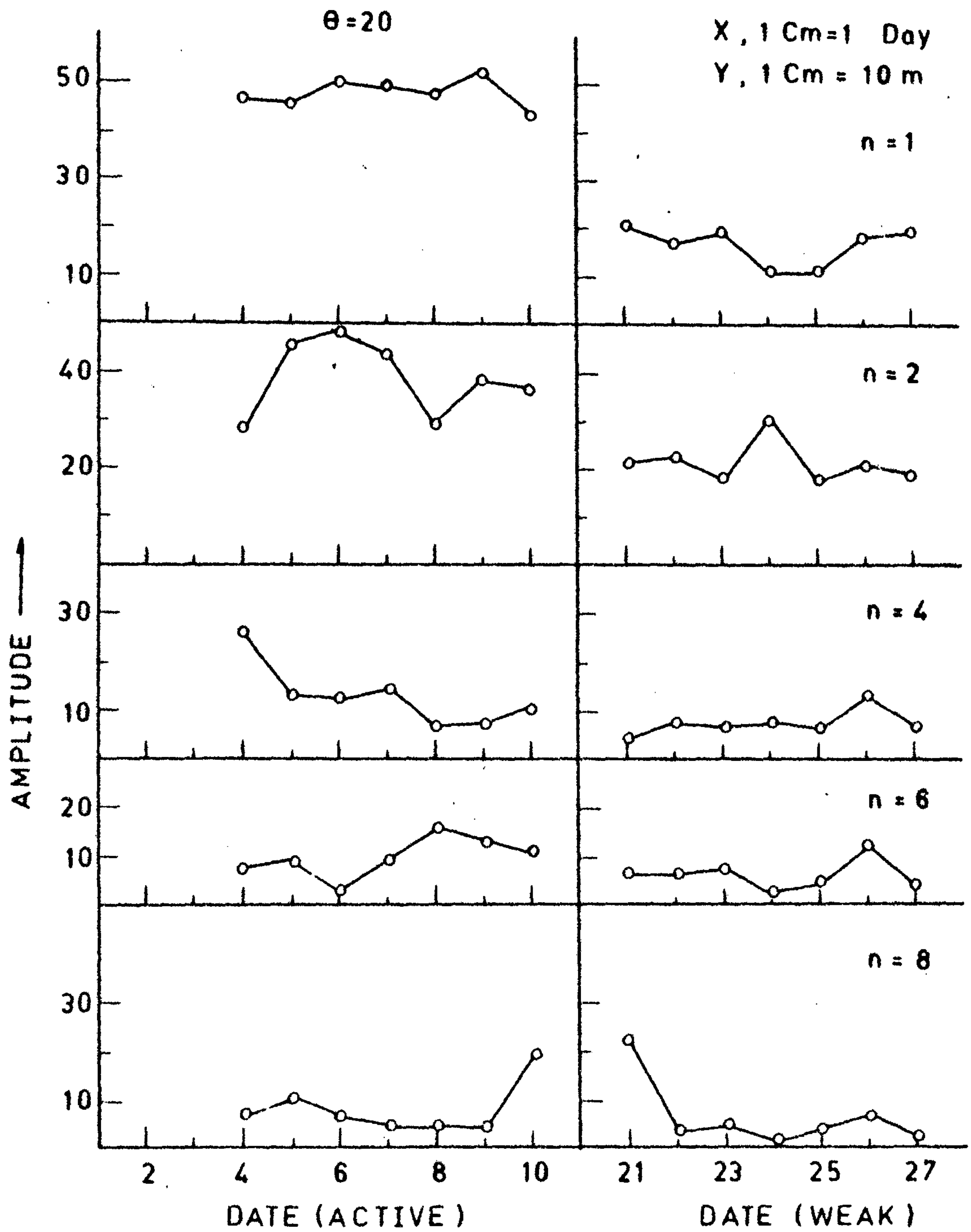


Fig.5.1(c). Variation of amplitudes of geopotential height field for wave number 1,2,4,6 and 8 during strong and weak monsoon phases.

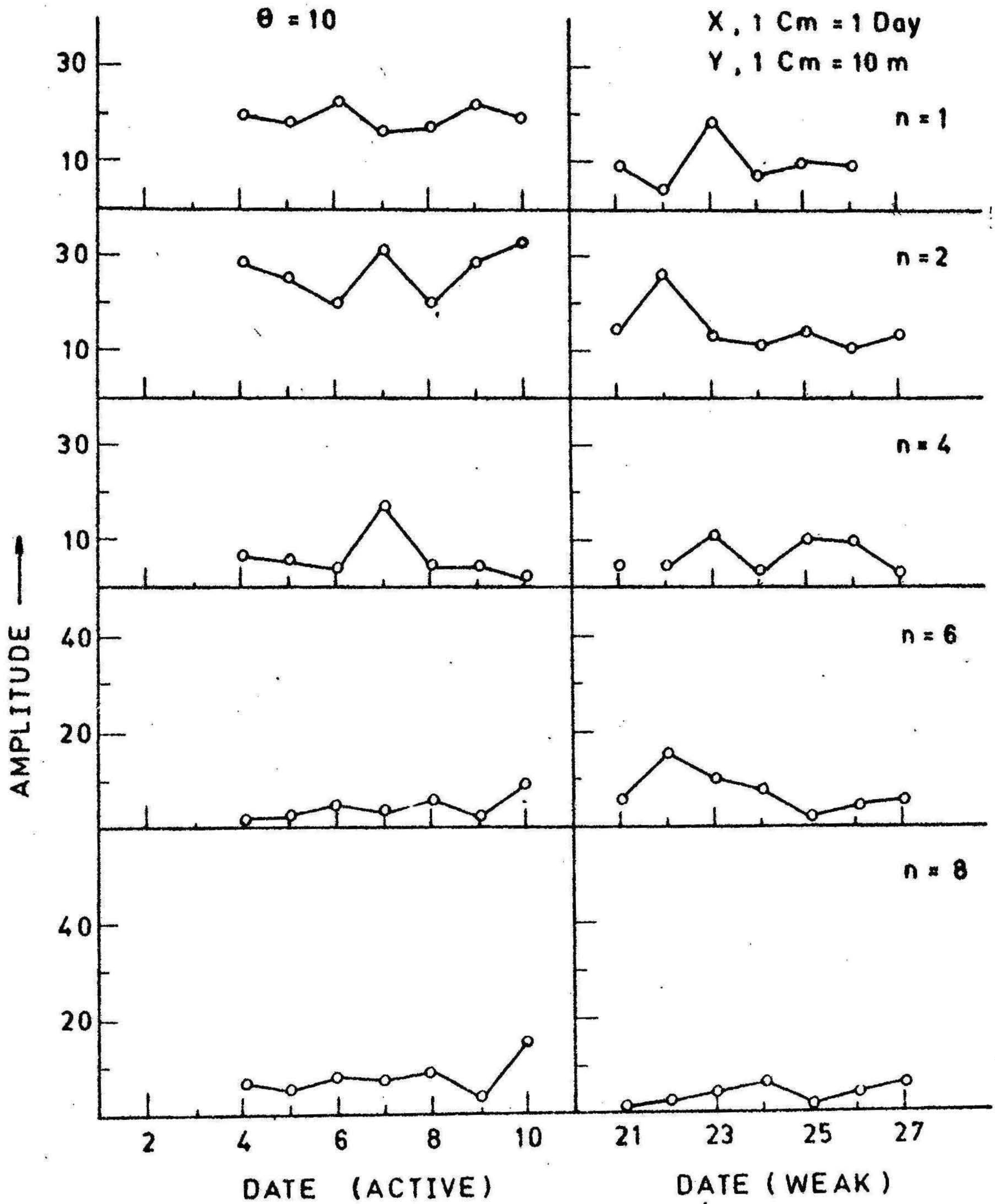


Fig. 5.1. (d). Variation of amplitudes of geopotential height field for wave number 1, 2, 4, 6 and 8 during strong and weak monsoon phases.

Figure 5.2(a) to (d) gives the distribution of mean kinetic energy for the two periods for the zonal flow and wave components 1 to 15. For the latitudes 10°N, 20°N, 30°N and 40°N, the kinetic energy of the zonal flow and the meridional flow as well as the total kinetic energy are shown in the figure.

From Figure 5.2(~~a~~ d) we see that at 10°N the kinetic energy contents for wave number 1 to 3 are more during the active period than during the weak period. Along latitude 20°N (Figure 5.2(~~b~~ c)) also the total kinetic energy contents for wave number 1 to 3 are more during the active period than the weak period. We observe a minor kinetic energy maximum around wave number 8 to 10 particularly during the active monsoon phase. The higher energy contents in the planatary scale waves in these figures also reflect the increased strength of planetary scale flow constituted by wave number 1 to 3 during the active monsoon period as compared to the weak monsoon period. The maximum around wave number 8 may be associated with transient active low latitude disturbances during the monsoon season.

The kinetic energy distribution between various waves for latitude 30°N (Figure 5.2(~~c~~ b)) is nearly similar for the two phases of the monsoon. This latitude is the transition latitude between the monsoon regime and the mid-latitudes where the effect of monsoon flow and the

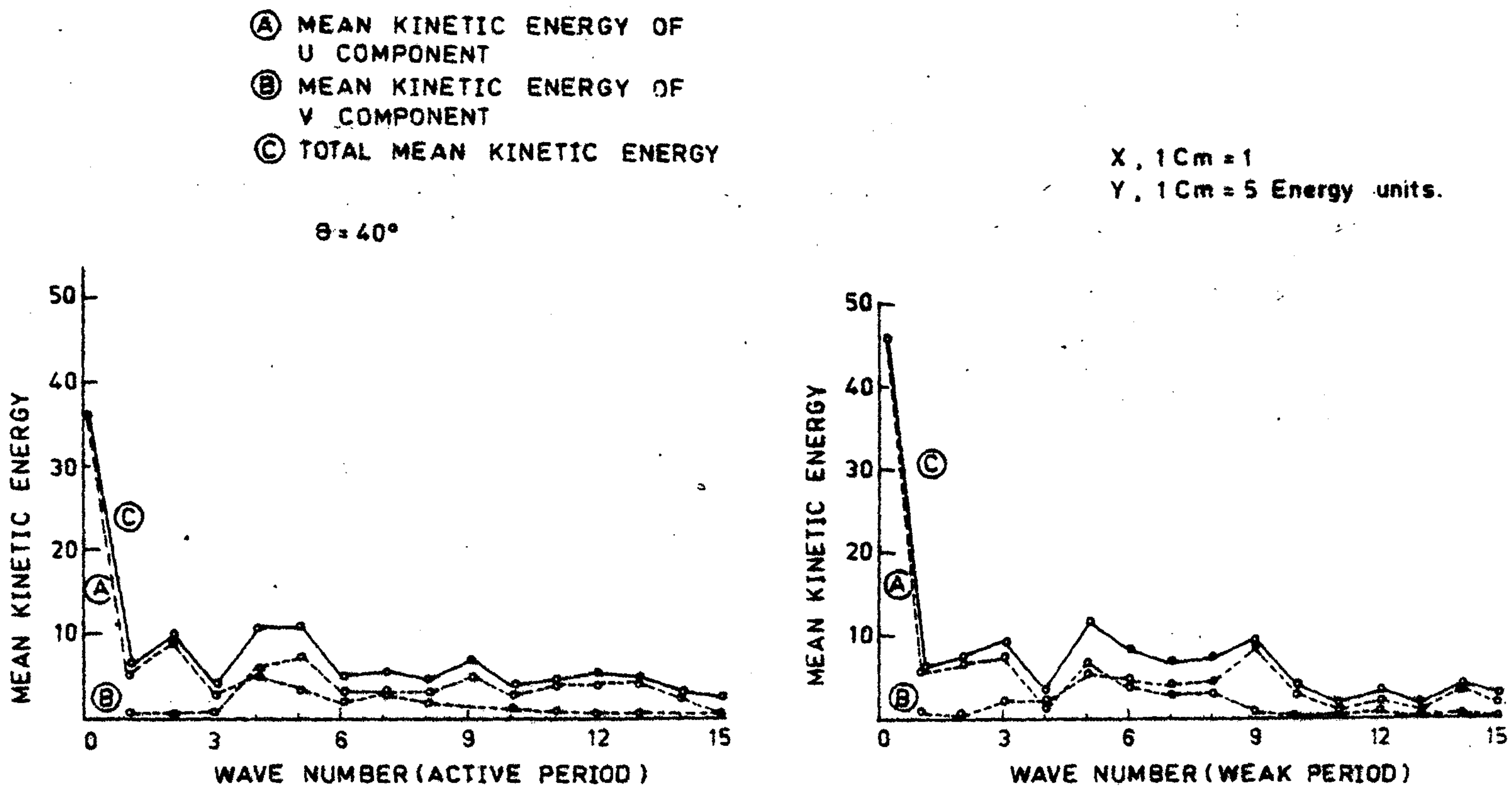


Fig. 5.2 (a) : The spectral distribution of kinetic energy during active and weak monsoon phases. Unit. $m^2\ sec^{-2}$

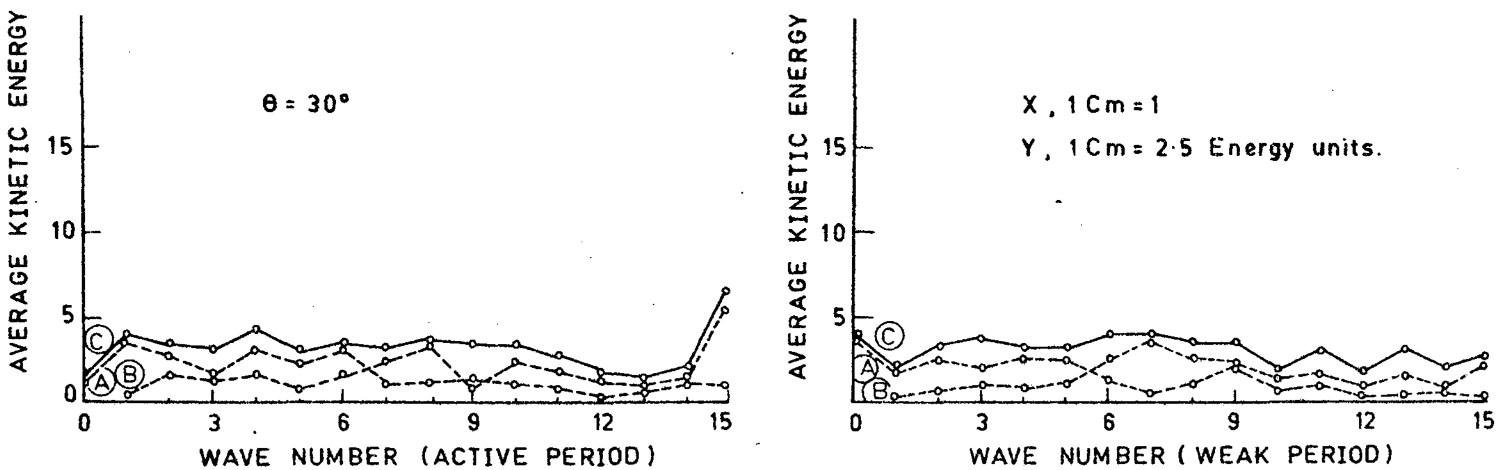


Fig. 5·2(b). The spectral distribution of kinetic energy during active and weak monsoon phases. Unit. m^2 sec^{-2}

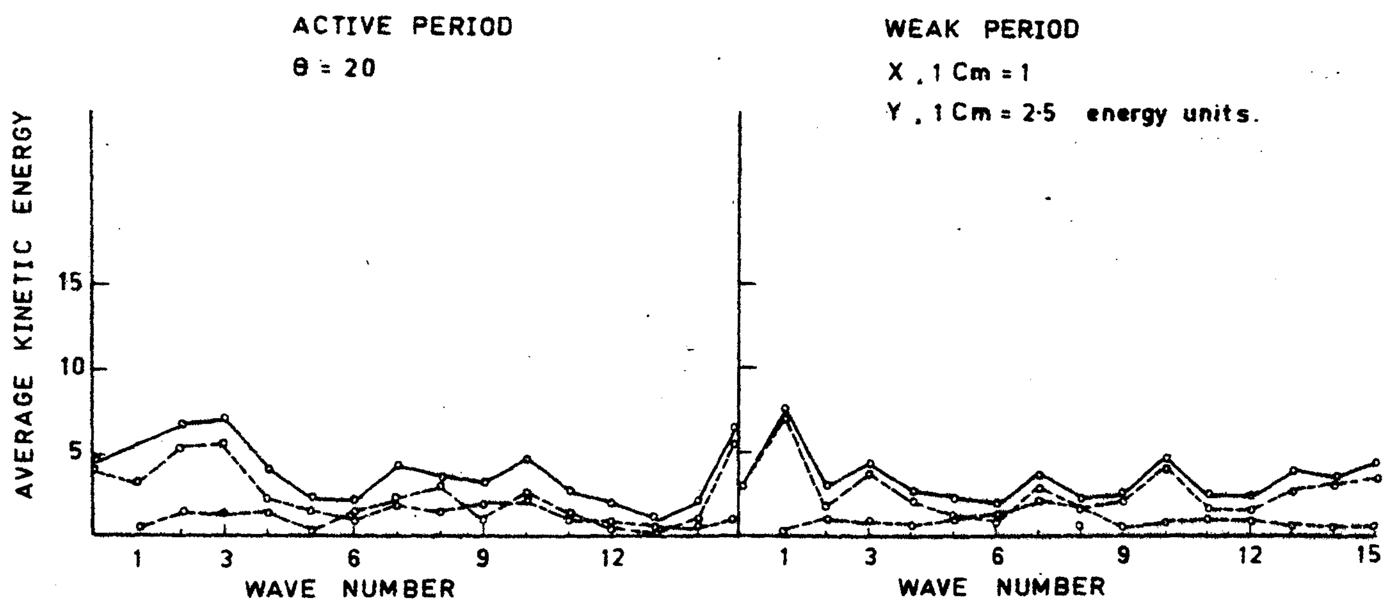


Fig. 5.2 (c). The spectral distribution of kinetic energy during active and weak monsoon phases. Unit. $m^2\ sec^{-2}$

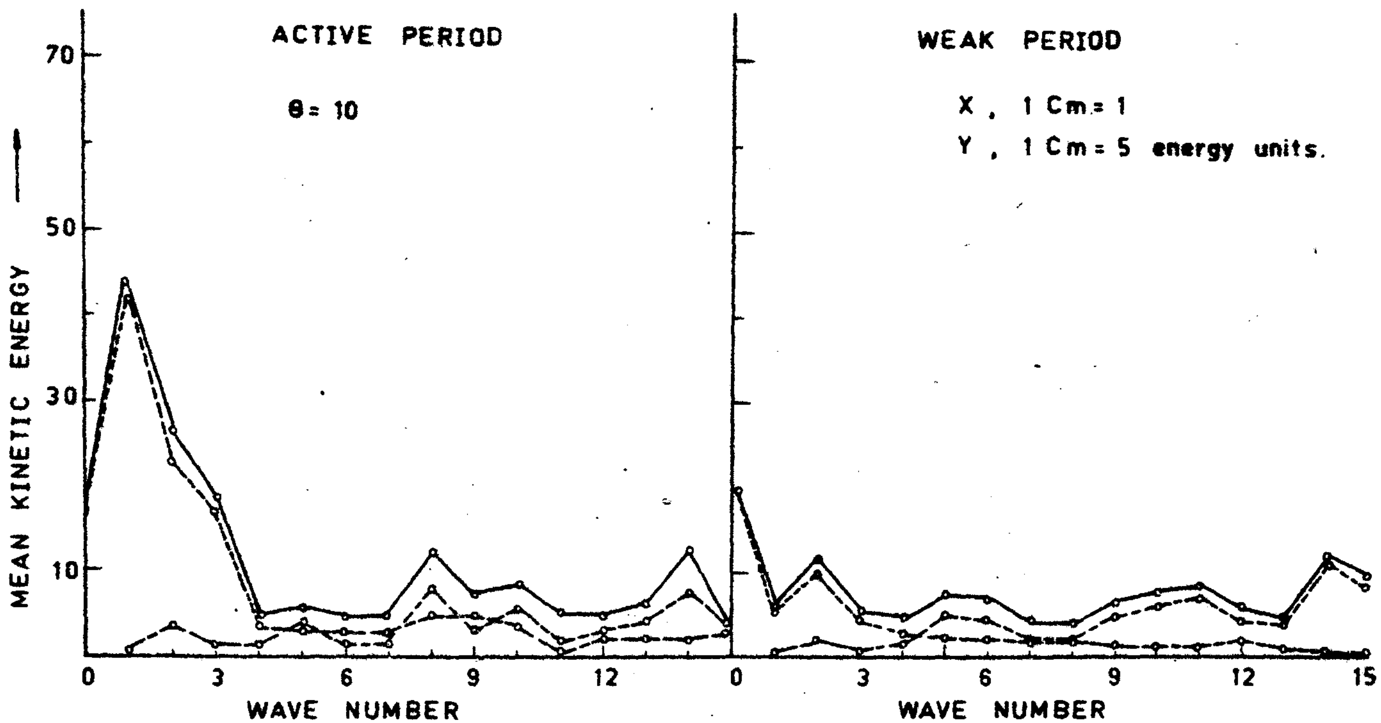


Fig. 5.2(d). The spectral distribution of kinetic energy during active and weak monsoon phases. Unit. $m^2 sec^{-2}$

westerly systems both appear to be weak.

As may be seen from Figure 5.2(d), the kinetic energy of zonally averaged flow at 40°N during the active monsoon period is less than that during the weak monsoon period. The weak zonal flow indicates low index circulation and is often believed to be associated with weak monsoon. In the present case the weak zonal index is associated with strong monsoon. The reason for this is not very clear. We also see that the kinetic energy of Rossby type waves from wave number 5 to 9 during the weak monsoon is greater than that during the active monsoon. This is consistent with the general concept that migratory mid-latitude disturbances are not favourable for an active monsoon.

SUMMARY

We have developed the method of Fourier analysis of geopotential height field along a latitude circle. Using this method, the spectral components of the zonal and meridional wind have been derived through geostrophic assumption and kinetic energy has been calculated. The comparison has been made of the amplitudes of geopotential height field and the kinetic energy during active and weak monsoon phases. Although the data samples are very small, the results reveal that the active monsoon over India on a regional scale is associated with the strong flow of planetary scale monsoon components of wave number 1 to 4. Thus, the Indian monsoon appears to be part of planetary scale tropical circulation during the summer season. On account of distribution of ocean and land over the area and trapping of moisture and heat by the Himalayan barriers and other mountains over the region, copious rain occurs over the Indian sub-continents. The conclusions derived above are based on a very limited set of data, and are only indicative but not conclusive. There is a need for study of atmospheric energetics in relation to the behaviour of summer monsoon utilising a large sample of data.

CHAPTER VI

SUMMARY

The thesis has attempted a brief survey of the general circulation of the atmosphere. The study has been divided into six chapters. The chief results of various chapters are summarised below.

1. Chapter I is devoted to the evolution of research on the understanding of the atmospheric general circulation from sixteenth century onwards. The basis for the general circulation as postulated by early scientists like Halley and Hadley have been discussed. Starting with Bjerknes, the studies by recent scientists in the field have been discussed. Finally brief details of various techniques and models which are useful to study the general circulation of the atmosphere are given.

seventeenth

2. Under chapter II, the equations for general circulation studies are discussed. The basic equations governing the atmospheric flow together with necessary assumptions have been considered. The equations related to the general circulation of the atmosphere in both (x,y,z) and (x,y,p) co-ordinate systems have been

discussed. Finally the procedure for decomposition of general circulation statistics into mean and eddies has been outlined.

3. Under chapter III, some observed features of atmospheric general circulation are discussed. The zonally averaged features of wind and temperature fields have been discussed together with their possible causes. Primary conclusions from this chapter are:

(i) both the westerly angular momentum and net heat input in the tropics is more than those in the higher latitudes.

(ii) The meridional transport of angular momentum and heat is accomplished both by the mean meridional circulation and by the eddy fluxes.

(iii) The westerly jet stream is maintained both by the convergence of angular momentum by eddies and by the effect of meridional temperature gradient through thermal wind equation.

It has been concluded that the earth-atmosphere system has developed an inbuilt mechanism by which the momentum and heat are transported from the excess to deficit regions and are simultaneously destroyed by frictional and radiative effects.

4. Under chapter IV, the energetics of the atmosphere have been discussed. It describes various topics of energies in the atmosphere, like kinetic energy, potential energy, internal energy and available potential energy. Finally the observed features of atmospheric energy cycle are discussed. The conclusions drawn from this chapter are:

(i) The total as well as zonal available potential energy is much larger in magnitude than the corresponding kinetic energy of the atmosphere. But, the eddy kinetic energy is larger in amount than the eddy available potential energy.

(ii) The energy cycle in the troposphere is such that the troposphere behaves like a heat engine. The zonal available potential energy is converted into eddy available energy which, in turn, converts into eddy kinetic energy and then to zonal kinetic energy.

(iii) The stratospheric energy cycle is such that the atmosphere works as a thermodynamic refrigerator. The source of energy for this region is the eddy kinetic energy and eddy available potential energy which have been found to be transported from the troposphere.

5. In the fifth chapter spectral distribution of kinetic energy over low and middle latitudes during an active and a weak monsoon phase has been discussed. The method of decomposing the geopotential height field into Fourier components and deriving the spectra of zonal and meridional velocities from them has been discussed. The amplitudes of geopotential height and kinetic energy of different wave components during the two phases have been examined. The results show that during an active monsoon phase the planetary scale waves consisting of wave number 1 to 4 contain more energy than during the weak monsoon phase.

The study reported in the thesis covers only a few aspects of the general circulation of the atmosphere. The kinetic energy calculations for the active and weak monsoon period are based on a very limited set of data, and the results require to be verified, by using a larger data set. In future, the author plans to extend the study using more data and also to cover other aspects of atmospheric energy processes like barotropic and baroclinic energy conversions, the generation of available potential energy and the dissipation of kinetic energy. The variation of various energy processes in

relation to the monsoon activity c
interesting and useful to understa
cts of summer monsoon circulation
weather .

REFERENCES

Bigelow, F.H. 1900: Report on the international cloud observations. Report of the chief of the weather beureau, 1898-99, Vol. II, Washington, 787 pp.

Bjerknes, V. 1937: Application of line integran theorems to the hydrodynamics of terrestrial and cosmic vortices. Astrophys. Norv., Vol. 2, No. 6, 263-339.

Bjerknes, V., Bjerknes, J., Bergeron, T. and Solberg, H. 1933: Physikalische Hydrodynamik. Berlin, J. Springer, 797 pp.

Buch, H. 1954: Hemispheric wind conditions during the year 1950. Final report, part 2, Contract AF 19(122)-153, Dept. of Meteorology, Mass. Inst. of Technology (32, 80).

Defant, A. 1921: On the dynamics of geostrophic winds. Q. J. Roy. Meteor. Soc., 52, 85-104.

Dove, H.W. 1835: Uber den Einfluss der 'Drehung der Erade auf die Stronungen ihrer Atmosphere. Poggerndorff's Ann. Phys. and Chem., 36, 321-351.

Faultz, D. 1951: Experimental analogies to atmospheric motions. Compendium of Meteorology, T.F. Malone, Ed. Boston, Amer. Meteor. Soc., 1235-1248.

Ferrel, W. 1856: An essay on the winds and currents of the ocean. Nashville J. Medicine and Surgery, 11, 287-301.

Hadley, G., 1735: Concerning the cause of general trade-winds. Phil. Trans., 29, 58-62.

Halley, E., 1686: An historical account of the trade-winds and monsoons abservable in the seas between and near the tropics with an attempt to assign the physical cause of said winds. Trans., 26, 254-267.

Helmholtz, H.V.: Uber atmospharische Bewegungen. Sitz-Ber, Akad. Wiss. Berlin, 647-663. English Trans. Abbe, C. 1893: The mechanics of the earth's atmosphere. Washington, Smithsonian Inst., 78-93.

Hildebrandsson, H.H. and Teissence de Bort, L. 1900: Les bases de la meteorologie dynamique. Vol. 2, Paris, Gauthier-Villars, 345 pp.

Lorenz, E.N. 1955: Available potential energy and the maintenance of the general circulation. Tellus, 7, 157-167.

Lorenz, E.N. 1960: Generation of available potential energy and the intensity of the general circulation. Dynamics of Climate. R.L. Pfeffer, Ed. New York, Pergammon Press, 86-92.

Maury, M.F. 1855: The physical geography of the sea. New York, Harper, 287 pp.

Newell, R.E., Kidson, J.W., Vincent, D.G. and Boer, G.J. (ed.) 1972: The general circulation of the Tropical atmosphere, Vol. 2, MIT Press, Cambridge, Mass.

Obasi, G.O.P. 1963: Atmospheric momentum and energy calculations for the southern hemisphere during the IGY. Sci. Report No. 6, Contract AF 19(604)-6108, Dept. of Meteorology, Mass. Inst. of Technology.

Oberbeck, A. 1888: Uber die Bewegungserschein ungen der Atmosphare. Sitz-Ber. Akad. Wiss. Berlin, 383-395, 1129-1138.

Oort, A.H. 1964a: On estimates of the atmospheric energy cycle. Mon. Weather Rev., 192, 483-493.

Oort, A.H. 1964b: On the energetics of the mean and eddy circulations in the lower stratosphere. Tellus, 16, 309-327.

Oort, A.H. and E.M. Rasonsson, 1971: Atmospheric circulation statistics, U.S. Dept. of Commerce, NOAA, 1971, p. 323.

Priestley, C.H.B. 1949: Heat transport and zonal stress between latitude. Q. J. Roy. Meteor. Soc., 75, 28-40.

Saltzman, B. 1957: Equations governing the energetics of the larger scales of atmospheric turbulence in the domain of wave number. J. Meteor. 14, 513-523.

Saltzman, B. 1958: Some hemispheric spectral statistics, J. Meteo., 15, 259-263.

Saltzman, B. and Fleisher, A. 1961: Further statistics on the modes of release of available potential energy. J. Geophys. Res., 66, 2271-2273.

Saltzman, B., Gottuso, R.M. and Fleisher, A. 1961: The meridional eddy transport of kinetic energy at 500 mb, Tellus, 13, 293-295.

Saltzman, B. and Teweles, S. 1964: Further statistics on the exchange of kinetic energy between harmonic components of atmospheric flow. Tellus, 16, 432-435.

Starr, V.P. 1948: An essay on the general circulation of the earth's atmosphere, J. Meteorol., 5, 39-43.

Starr, V.P. 1954: Balance requirements of the general circulation. Geophys. Res. Pap. No. 35, 57 pp.

Starr, V.P. and White, R.M. 1951: A hemispherical study of the atmospheric angular momentum balance. Q. J. Roy. Meteor. Soc., 77, 215-225.

Starr,V.P. and White, R.M. 1953: A scheme for the study of hemispheric exchange processes. Q. J. Roy. Meteor. Soc. 337, 551-553.

Van Mieghem, J. 1952: Energy conversions in the general circulation. Tellus, 4, 334-351.

Wiin-Nielsen, A. 1962: On transformation of kinetic energy between the vertical shear flow and the vertical mean flow in the atmosphere. Mon. Weather Rev., 90, 311-323.

Wiin-Nielsen, A., Brown, J.A. and Drake, M. 1964: Further studies of energy exchange between the zonal flow and the eddies, Tellus, 16, 168-180.

White, R.M. 1951: The meridional eddy flux of energy Q. J. Roy. Meteor. Soc., 332.

Addendum

Houghton, H.G. 1954: J. Meteor. 11, pp. 223.

Krishnamurti, T.N. 1978: Monsoon Dynamics.
Birkhauser Verlog, ed., pp. 1529.

Riechl, H. and Yeh, T.C. 1954: Am. Meteo. Soc.
Meteor. Monograph. Vol. 2, No.7,

TABLE A: LIST OF SYMBOLS

a	=	Mean radius of earth, 6.37 x 10^3 Km
c_p	=	Specific heat of air at constant pressure, 9.96 x 10^6 $Cm^2Sec^{-2}deg^{-1}$
c_v	=	Specific heat of air at constant volume, 7.09 x 10^6 $Cm^2Sec^{-2}deg^{-1}$
F_z ? F_x, F_y	=	Components of Frictional Force in the X and Y directions respectively.
f	=	Coriolis parameter, 2 sin
g	=	Acceleration of gravity, 9.8 m/sec^{-2}
$\vec{k}$	=	Unit vector directed upward
p	=	Pressure
p_{00}	=	Standard pressure, 1000 mb
$\dot{Q}$	=	Rate of heating per unit mass
R	=	Gas constant for air, 2.87 x 10^6 $Cm^2Sec^{-2}deg^{-1}$
T	=	Absolute temperature
t	=	Time
u	=	Eastward Component of in the X-direction
$\vec{V}$	=	Wind velocity
v	=	Northward component of in the Y-direction
ω	=	Upward Component of in the Z-direction

α = Specific volume

Γ = Vertical lapse rate of temperature

Γ_d = Dry-adiabatic lapse-rate, 9.8°/Km

θ_T = Potential temperature

θ = Latitude

λ = Longitude

ρ = Density

K = R/c_p , 0.288

ϕ = Geopotential, gz

Ω = Angular velocity of the earth, 7.292×10^{-5} Sec^{-1}

$\omega = \frac{dp}{dt}$, vertical p-velocity.