STRUCTURE AND VARIATIONS OF MONSOON FLOW

T.N. Krishnamurti
Department of Meteorology, Florida State University
Tallahassee, Florida, U.S.A.

Summary: In this paper we discuss the basic workings of the monsoon system; i.e. how differential heating drives the monsoon. Two major papers – one from the studies of the late Adrian Gill dealing with the response of the atmosphere to antisymmetric heating and the other by Murakami, Godbole and Kelkar on a zonally symmetric monsoon – are reviewed.

1. INTRODUCTION

Large scale differential heating between land and ocean is a primary forcing of the monsoon. Fig. (1) illustrates the vertically integrated heating field (in units of °K/day). Solid lines denote heating and dashed lines denote net cooling. This illustration is based on a recent study, Johnson et al. (1987). The two panels figs. 1a and 1b illustrate the differential heating for the winter and the summer months. Normally a net heating over Indonesia and the equatorial western Pacific ocean, and a net cooling over Northern China and Siberia constitutes this differential heating during the winter months, fig. 1a. The picture during the northern summer months shows an axis of net heating near 20°N extending eastwards from the Northern Bay of Bengal to the Indochina peninsula. The net cooling is along 30°S which extends from the Mascarene Islands to Western Australia. This field of differential heating is an important part of the monsoon system. A principal axis of the annual cycle of the monsoon may be defined following the region of the maximum monthly mean rainfall. That rainfall belt migrates from Indonesia to the foothills of the Himalayas between January and August and makes its return traverse between September and December. This axis exhibits considerable interannual variability in its position and intensity. During the Elnino years, an eastward and equatorward shift of the net heating region is a major part of the interannual variability.

Fig. 2 illustrates the normal evolution of the annual cycle of the monsoon. Here we show a schematic of the vertical structure of some of principal elements. Among these is a prominent low level counter clockwise gyre which migrates northward between January and August that is more clearly portrayed in fig. 3 from a study of Findlater (1971). Here the clockwise gyre of the low level wind maximum is shown to move northwards with season. This gyre terminates in the monsoon trough where the maximum rainfall is found.

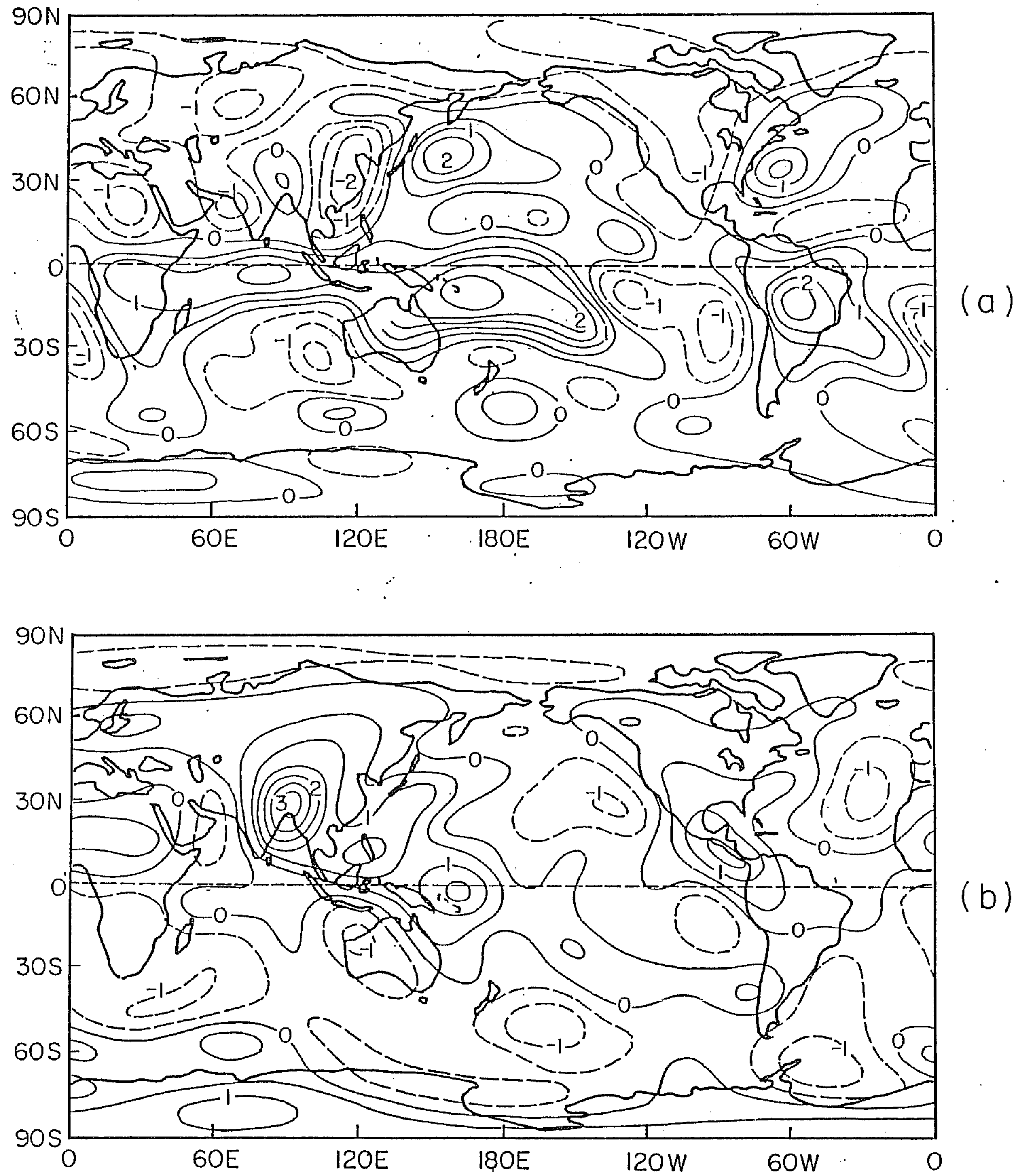


Fig. 1. Vertically integrated net diabatic heating for a) January b) July. Units °C/day; solid lives heating, dashed lives cooling. Based on Johnson et.al. (1987).

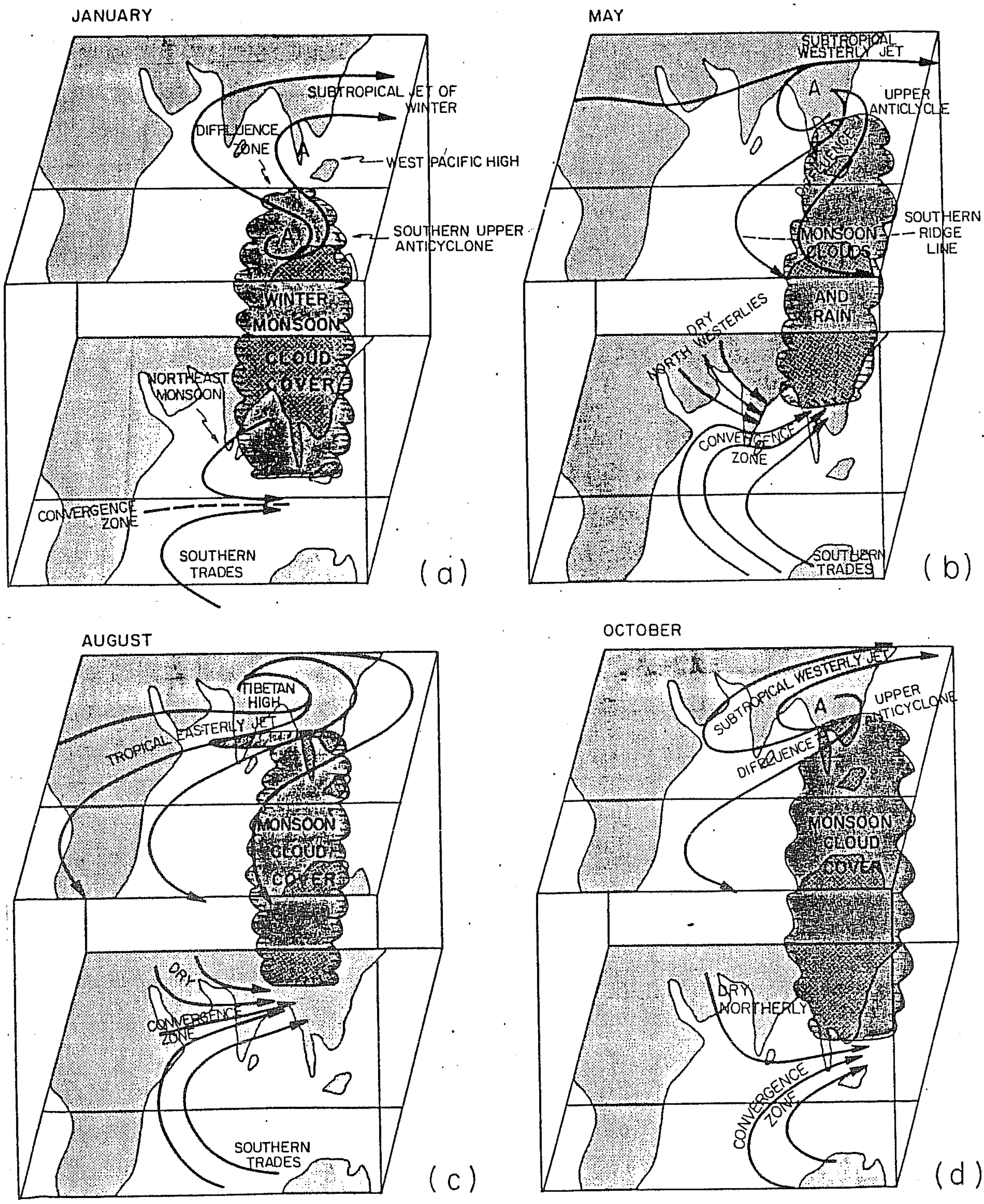


Fig. 2. Schematic monsoon circulatia during a) January, b) May, c) August and d) October.

The northward motion of the gyre is closely related to the motion of the maximum monsoon rainfall along its principal axis. The ascending air over this rainfall belt diverges in the upper troposphere. The weather system above the precipitation belt in the upper troposphere is an anticyclone except when precipitation occurs over the equatorial latitudes in which case two anticyclones, one on either side of the equator, are found. This split is related to a lack of geostrophic adjustment of the divergent flow over the equatorial latitudes. Fig. 2 illustrates the three dimensional structure of the monsoonal inflow and outflow from the rain belt. The north–south migration of the monsoon system is controlled by the evolution of the differential heating. During the northern winter (summer) the cooling (heating) of the land masses appear to provide the major forcing. The oceanic regions supply the needed moisture. Latent heating is the major component of heating; deep tropospheric cumolonimbrus heating is important for driving the large scale monsoon circulation shown in fig. 2. The elements of the monsoon circulations for the winter and the summer monsoon are further highlighted in fig. 4. There are basic similarities in the winter and the summer monsoon system. The Siberian high of the winter component has somewhat of an analogous role to that of the Mascarene high of the summer monsoon. The former is located over a land mass while the latter is found mostly over the southwestern Indian Ocean. The cross equatorial and the Somali jet of the Northern summer flows is paralleled by the low level North east monsoon flow and the strong winter surges along the eastern coasts of Asia. The monsoon trough are regions of heavy rain in both systems. A warm troposphere extends vertically above the monsoon trough in both systems. In the upper troposphere the Tibetan high pressure cell of summer monsoon has its counterpart in the West Pacific upper Anticyclone of the winter monsoon. The upper tropospheric monsoon circulations include two jet streams, the tropical easterly jet on the southern flank of the Tibetan high near 10°N and 150 mb during the summer and the subtropical westerly jet on the northern flank of the West Pacific anticyclone near 30°N and 200 mb during the Winter.

For the basic understanding of the monsoon system it is important that any simple theory or model have the capability to describe some of the above salient elements of this system.

2. THE ADRIAN GILL THEORY OF THE MONSOON SYSTEM

Perhaps our lowest order understanding of the monsoon system is best provided by a simple theory of Adrian Gill (1981). In his text on Atmosphere ocean dynamics, Gill (1984), he has developed a theory for the response of the atmosphere to simple heating distributions.

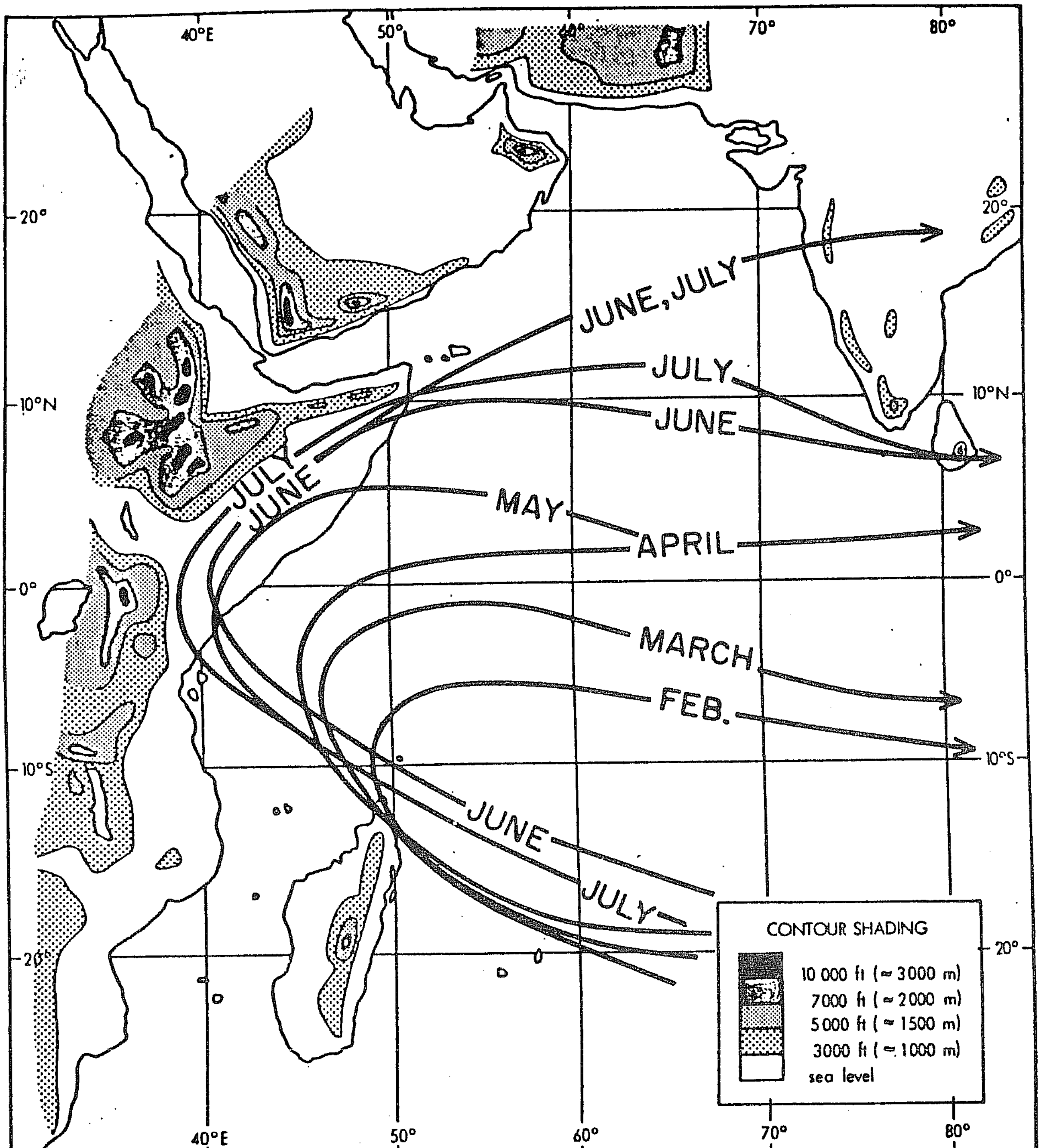


Fig. 3. Monthly mean axis of low level jet. Findlater (1971).

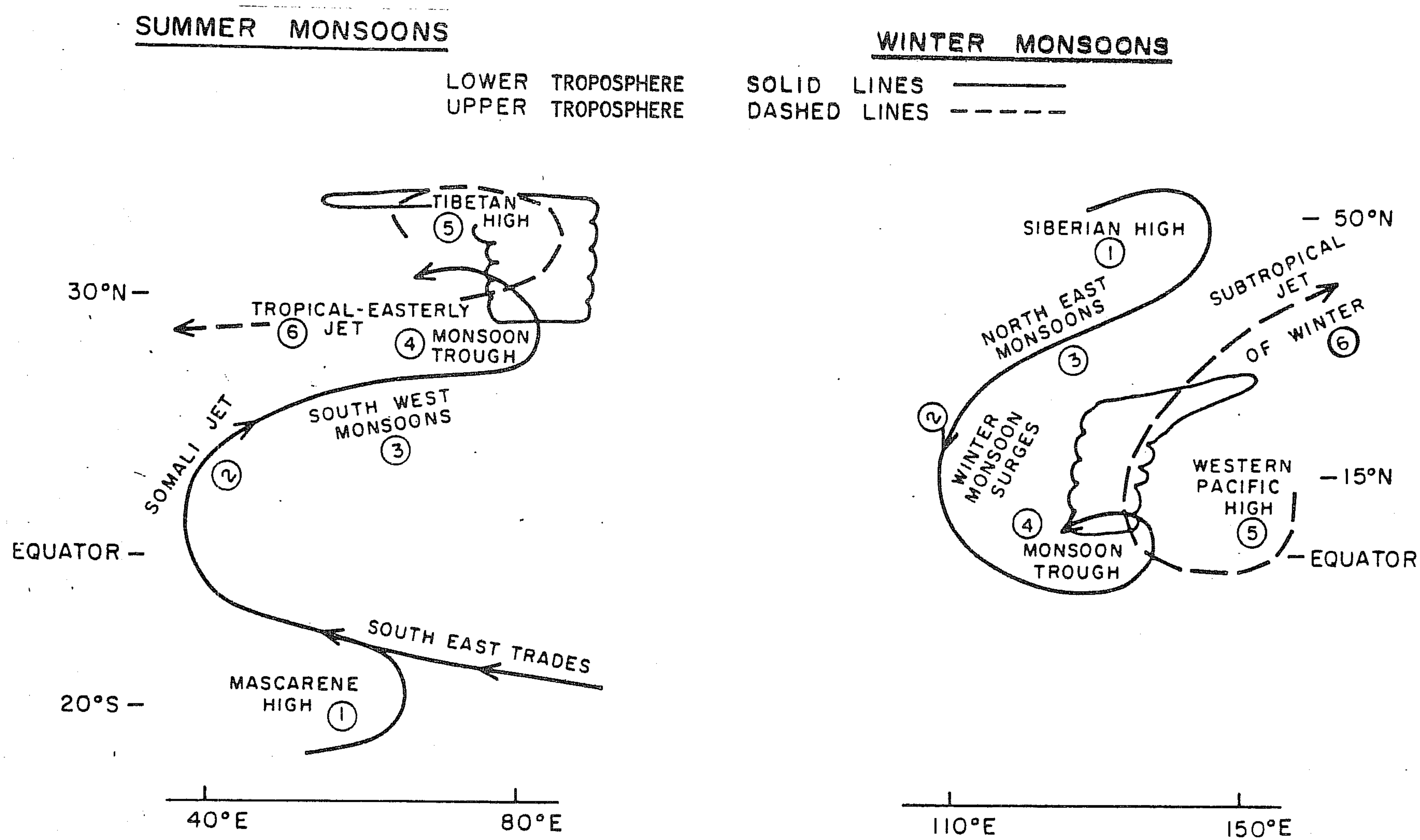


Fig. 4. Elements of the summer and the winter monsoon systems.

This is a simple linear, analytic model which elucidate some basic features of the response of the tropical atmosphere to diabatic heating concentrated in an area of finite extent.

One assumes that the basic state is at rest upon which small perturbations are superimposed. Since the motion is confined to the tropics, the equatorial ß–plane approximation is used. Using the equatorial Rossby radius $(c/2\beta)^{\frac{1}{2}}$ as a length scale which is about 10° of latitude when the equivalent depth is 400 m. and the time scale $(2\beta c)^{-\frac{1}{2}}$ which is about 6 hours, the nondimensional form of the governing equations are:
Equation of motion,

$$U_t - \tfrac{1}{2}yv = -P_x \tag{2.1}$$

$$V_t + \tfrac{1}{2}yu = -P_y \tag{2.2}$$

Mass continuity equation,

$$P_t + U_x + V_y = -Q \tag{2.3}$$

where Q denotes the diabatic heating it is in fact a diabatic mass sink in the mass continuity equation. Other valuables have the usual meanings. Note that (2.1)–(2.3) describes a forced shallow water system. From (2.3), the vertical velocity is given by

$$W = P_t + Q \tag{2.4}$$

Now introducing 'Rayleigh friction' or 'Newtonian cooling' to the system, replacing $\partial/\partial t$ by $\partial/\partial t + \varepsilon$ (here ε is a dissipative parameter), the response of steady state system to the steady forcing is then described by ε

$$\varepsilon u - \tfrac{1}{2}yv = -P_x \tag{2.5}$$

$$\varepsilon v + \tfrac{1}{2}yu = -P_y \tag{2.6}$$

$$\varepsilon p + u_x + v_y = -Q \tag{2.7}$$

$$W = \varepsilon_p + Q \tag{2.8}$$

Now in this system, Q and ε can be viewed as energy 'source' and energy 'sink', respectively. Because of the existence of both Q and ε, the equilibrium state of the system under the forcing Q and dissipating ε can be realized. From scale analysis, assuming that Q has a y–scale of order unity and ε is small, eq. (2.6) becomes

$$\tfrac{1}{2}yu = -P_y \tag{2.9}$$

This indicates that the zonal flow is in geostrophic balance. Now (2.5), (2.7), (2.8) and (2.9) constitute a linear system for U, V, P and W after specification of Q and ε; whereas W is determined by P and Q. Thus the linear system is constituted by (2.5), (2.7) and (2.9).

The solution of the system is obtained as follows:

(1) Defining new variables

$$q = p + u \tag{2.10}$$

$$\gamma = p - u, \tag{2.11}$$

then (2.5), (2.7) & (2.9) lead to

$$\varepsilon q + q_x + v_y - \tfrac{1}{2}yv = -Q \tag{2.12}$$

$$\varepsilon\gamma - \gamma_x + v_y + \tfrac{1}{2}yv = -Q \tag{2.13}$$

$$q_y + \gamma_y + \tfrac{1}{2}yq - \tfrac{1}{2}y\gamma = 0 \tag{2.14}$$

the solution of q, γ and v in (2.12)–(2.14) have the form of parabolic cylinder functions $D_n(y)$ when Q=0. The forced problems can thus be solved by expanding q, γ, v and Q in terms of these functions, i.e.,

$$\begin{bmatrix} q \\ \gamma \\ v \\ Q \end{bmatrix} = \sum_{n=0}^{\infty} \begin{bmatrix} q_n(x) \\ \gamma_n(x) \\ v_n(x) \\ Q_n(x) \end{bmatrix} D_n(y) \tag{2.15}$$

With the aid of the following properties

$$(D_n)_y + \tfrac{1}{2}yD_n = nD_{n-1} \tag{2.16}$$

$$(D_n)_y - \tfrac{1}{2}yD_n = -D_{n+1}, \tag{2.17}$$

Eqs. (2.12)–(2.14) can be written after substituting from (2.15),

$$\varepsilon q_o + \frac{dq_o}{dx} = -Q_o \tag{2.18}$$

$$\varepsilon q_{n+1} + \frac{dq_{n+1}}{dx} - v_n = -Q_{n+1} \quad n \geq 0$$

$$\varepsilon\gamma_{n-1} - \frac{d\gamma_{n-1}}{dx} + nv_n = -Q_{n-1} \quad n \geq 1 \tag{2.19}$$

$$q_1 = 0$$

$$\gamma_{n-1} = (n+1)q_{n+1} \quad n \geq 1 \tag{2.20}$$

(2) Consider 2 special cases of heating.

a) Q is symmetric about the equator

$$Q(x,y) = F(x)D_o(y) = F(x)\exp(-\tfrac{1}{4}y^2) \tag{2.21}$$

b) Q is antisymmetric about the equator

$$Q(x,y) = F(x)\,D_1(y) = F(x)y\exp(-\tfrac{1}{4}y^2)$$

Note that

$$(D0, D_1, D_2, D_3) = (1, y, y^2-1, y^3-3y)\exp(-\tfrac{1}{4}y^2) \tag{2.22}$$

and in both cases,

$$F(x) = \begin{cases} \cos kx & |x| < L \\ 0 & |x| > L \end{cases} \qquad (2.23)$$

where $k = \pi/2L$. Thus, heating is confined in a channel bounded by $x = \pm L$.

The symmetric heating is more relevant for a description of the winter monsoon while the antisymmetric heating provides a reasonable model for the summer monsoon.

(3) The following is a summary of the response for both cases.

Table 1

Symmetric case	Asymmetric case
Response Part 1.	*Response Part 1*
Symmetric case ($Q_o = F(x)$)	Asymmetric case ($Q1 = F(x)$)
n=0, q_o only	n=0, q_1 only
damped Kelvin wave	mixed planetary gravity wave
Eastward propagating	do not propagate
phase speed = 1	zero solution for $x>\|L\|$,
decay rate = ε	for $x<\|L\|$, then
zero solution for $x<-L$	$q_1=0$,
for $x > -L$:	$v_o=Q_1$
$u=P=\frac{1}{2}q_o(x)\exp(-\frac{1}{4}y^2)$	
$v=0$	
$w=\frac{1}{2}\{\varepsilon q_o(x) + F(x)\}\exp(-\frac{1}{4}y^2)$	

This represents Walker circulation

Symmetric case	Asymmetric case
Response Part 2.	*Response Part 2.*
n=1, q_2 only	n=2, q_3 only
long planetary wave	long planetary wave
westward propagating	westward propagating
phase speed = 1/3	phase speed = 1/5
decay rate = 3ε	decay rate = 5ε
zero solution for x>L	zero solution for x>L
for x<L, then	for x<L, then
$p=\frac{1}{2}q_2(x)(1+y^2)\exp(-\frac{1}{4}y^2)$	$p=\frac{1}{2}q_3(x)y^3 \exp(-\frac{1}{4}y^2)$
$u=\frac{1}{2}q_2(x)(y^2-3)\exp(-\frac{1}{4}y^2)$	$u=\frac{1}{2}q_3(x)(y^3-6y)\exp(-\frac{1}{4}y^2)$
$v=\{F(x)+4\varepsilon q_2(x)\}y \exp(-\frac{1}{4}y^2)$	$v=\{6\varepsilon q3(x)(y2-1)+F(x)y2\}\exp(-\frac{1}{4}y2)$
$w=\frac{1}{2}\{F(x)+\varepsilon q_2(x)(1+y^2)\}\exp(-\frac{1}{4}y^2)$	
$w=\{\frac{1}{2}\varepsilon q_3(x)y^3+F(x)y\}\exp(-\frac{1}{4}y^2)$	

3. ZONALLY SYMMETRIC GENERAL CIRCULATION MODEL

Back in 1970, Murakami et al. (1970), presented a zonally symmetric general circulation model of a monsoon that simulated most of the elements of the circulation driven by differential heating between land and ocean. This model is extremely illustrative of the workings of the differentially heated monsoon system and its energetics. They integrated a set of zonally symmetric primitive equations using the earth's surface as a coordinate surface. The earth's surface was separated at 10°N between ocean to the south and land to the north. Smoothed mountains were included north of 25°N. The momentum, mass continuity, thermal and moisture balance equations were integrated in a meridional vertical plane between the south and the north pole. The atmosphere started from a state of rest and was driven by differential heating between land and ocean. The components of heating that are important for the monsoon are:

i) The heat balance of the land area, including the elevated Tibetan Plateau, viz.:
   (a) sensible heat flux from the land surface
   (b) latent heat flux from the land surface
   (c) incoming minus outgoing (or reflected) shortwave radiation
   (d) incoming minus outgoing longwave radiation
ii) Air/Sea interaction, i.e., fluxes of latent and sensible heat from the ocean
iii) Heating with the atmosphere which includes warming (or cooling) by the following processes
   (a) deep convection
   (b) shallow moist and dry convection
   (c) longwave radiation
   (d) shortwave radiation
   (e) adiabatic processes

Among the above processes, the most important are: deep convection, all of the elements of heat balance of the earth's surface, evaporative fluxes from the ocean, and the adiabatic warming and cooling within the atmosphere.

With the above framework Murakami et al. (1970) simulated a very realistic picture of the monsoons around 80°E longitude. Their simulations included the following features:

(a) realistic southeast trades south of the equator with maximum speed near 900 mb
(b) southwest monsoons over India with a maximum speed near 600 mb
(c) a tropical easterly jet in the upper troposphere near 150 km and around 10°N
(d) monsoon troughs (in the sea–level pressure field at 25°N)
(e) a Tibetan high at 30°N and 200 mb
(f) a warm troposphere between the monsoon trough and the Tibetan high
(g) a rainfall maximum at around 20°N
(h) a Hadley cell with rising motion around 20°N and descent near 20°S with

northeasterlies in the upper troposphere.

Much of what they simulated falls within the known range of observational variability of these well–known phenomena. The major drawback of this study is the absence of interactions of planetary–scale monsoon waves with the zonal flows and shorter scale waves.

A number of sensitivity studies were also conducted by Murakami et al. (1970). The purpose of these sensitivity studies was to investigate what were some of the important (and also less important) parameters in the successful simulation of monsoons. They found that the Himalayan mountains were very important as well as the salient physical processes listed above. Surprisingly, they noted that the intensity of monsoons and of the differential heating was not changed much if the sea–surface temperatures were slightly altered. The ocean was found to important for its supply of latent heat and the land for its heat balance and its warm temperatures. This framework is viewed as follows. The solar heating warms the land surface, the gradual increase of soil temperature results in super–adiabatic lapse rates and this is continuously removed in the model by dry convective adjustment over a substantial portion of the meridional belt. This results in the gradual formation and intensification of a heat low. Soon moist air from the ocean starts to converge into the general region of the heat low. Conditional instability over the land area grows and soon thereafter moist convection starts. Rainfall; warming of the troposphere; and the establishment of a Tibetan high, a monsoon trough and a Hadley cell, follow. The upper level tropical easterly jet forms due to a transformation of kinetic energy of meridional motions into zonal motions. In many ways the Murakami et al. framework of broadscale monsoons is akin to a giant sea breeze whose scale is of the order of 6000 km and in which the effects of the earth's rotation on the motions on the meridional plane become important.

An important relation between differential heating and the maintenance of strong circulations can be deduced from simple energy considerations. That relation simply states that in a closed system, if the rotational, divergent and available potential energies are increasing (or are in a statistical steady state), the maintenance of such a system against frictional dissipation requires a differential heating mechanism. The first application of this principle appeared in this zonally symmetric monsoon model. A simple zonally symmetric system may be described by the following equations. A list of symbols is provided in Table 2.

Table 2

List of useful symbols

u	zonal velocity
v	meridional velocity
ω	vertical velocity
θ	potential temperature
q	specific humidity
ψ	streamfunctions
χ	velocity potential
$A \cdot B$	energy exchange from a to B
K, K_u, K_v, K_ψ, K_x	kinetic energy
F_x, F_y	frictional force per unit mass of air

Momentum equations

$$\frac{\partial u}{\partial t} = -v\frac{\partial u}{\partial y} - \omega\frac{\partial u}{\partial p} + fv + F_x \tag{3.1}$$

$$\frac{\partial v}{\partial t} = -v\frac{\partial v}{\partial y} - \omega\frac{\partial v}{\partial p} - fu - g\frac{\partial z}{\partial y} + F_y \tag{3.2}$$

Hydrostatic relation

$$g\frac{\partial z}{\partial p} = -\frac{RT}{p} \tag{3.3}$$

Mass continuity equation

$$\frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} = 0 \tag{3.4}$$

First Level of Thermodynamics

$$\frac{\partial \theta}{\partial t} = -v\frac{\partial \theta}{\partial y} - \omega\frac{\partial \theta}{\partial p} + \frac{1}{C_p}\left[\frac{p_0}{p}\right]^{R/C_p} H \tag{3.5}$$

Here H denotes diabatic heating.

The energetics of this system may be expressed by the relations:

$$\frac{\partial}{\partial t} K_u = (K_v \cdot K_u) - D_x \tag{3.6}$$

$$\frac{\partial}{\partial t} K_v = (P \cdot K_v) - (K_v \cdot K_u) - D_y \tag{3.7}$$

$$\frac{\partial}{\partial t} P = -(P \cdot K_v) + G - D_p \qquad (3.8)$$

where K_u, K_v and P denote the zonal, meridional kinetic and available potential energy respectively, over the entire mass of the atmosphere and D_x, D_y and D_p denote dissipation of the three aforementioned energy quantities.

In a system that is either exhibiting an increase of K_u, K_v and P or is in a statistical steady state the following inferences can be made. If the dissipation terms D_x, D_y and D_p are positive definite quantities, then (i) $(K_v \cdot K_u)$ must be positive, (ii) $(P \cdot K_v)$ must be positive and (iii) a net generation (i.e., $G > 0$) is required to maintain this system as stated above. The generation term here is simply measured from a covariance of the diabatic heating and temperature.

Murakami et al. (1970) utilized such a framework to simulate a zonally symmetric monsoon in a general circulation model. The model included oceans to the south and land areas with mountains to the north. Other features were air–sea interaction, convective adjustment, detailed radiative processes and large scale condensation. With the inclusion of mountains and a mean July solar insolation input, the model simulated a realistic monsoon including such features as: the monsoon trough, monsoon rainfall, warm troposphere, tropical easterly jet, strong Hadley circulation and lower tropospheric monsoon westerlies.

A similar generalization on the role of differential heating and monsoon circulation for a fully three–dimensional motion was presented by Krishnamurti and Ramanathan (1982). The analogous energetics of a fully three–dimensional system may be written for the rotational kinetic energy K_ψ, divergent kinetic energy K_x, and available potential energy P over a closed region:

$$\frac{\partial K_\psi}{\partial t} = (K_x \cdot K_\psi) - D\psi$$

$$\frac{\partial K_x}{\partial t} = (K_x \cdot K_\psi) + (P \cdot K_x) - D_x$$

$$\frac{\partial P}{\partial t} = -(P \cdot K_x) + G - D_p$$

Here G denotes the generation term and dissipation terms are denoted by D_ψ, D_x, and D_p, respectively. It is worth noting that the pressure interactions are absent in the zonal energy equation of a zonally symmetric system since $\partial p/\partial x = 0$. When the fully three dimensional system is cast in terms of the rotational and divergent motions, pressure interactions again vanish from the rotational kinetic energy equation. The energetics of the two systems and their interpretations are quite analogous. That is not too surprising since the meridional wind in a zonally symmetric system is entirely divergent and the zonal wind is entirely rotational. Thus, in a situation such as the onset period of the monsoon when the rotational, divergent and available potential energies of the monsoon region are increasing with time, a net differential heating is required to generate available potential energy. The available potential energy must transfer energy to the divergent motions. Finally, we can make the statement that divergent motions must transfer energy to the rotational motions. All of the aforementioned arguments pre–suppose that we are dealing with a closed system. Fig. 7 presents an outline of the basic energetics of the zonally symmetric monsoon following Murakami et al. (1970).

4. DISCUSSION

For the symmetric heating, both the perturbation velocity field and the pressure field are also symmetric about the equator (Fig. 5 a,c). In the heating region, the flow field is dominated by the upward motion. Easterlies from east of the forcing region and westerlies from west of the forcing region move toward the heating center in the lower layer along the equator (Fig. 5 a,b). Outside the forcing, the subsidence is found everywhere (Fig. 5 c). The vertical circulation over the Pacific characterizes the Walker circulation. On the other hand, the perturbation pressure field is found negative everywhere with a trough at the equator in the easterly regime to the east of the forcing region and two low centers located on the north–west and south–west side of the forcing region (Fig. 5 b). The zonally integrated solution shows that on both sides of equator lower layer is dominated by easterlies & higher layer is dominated by westerlies. Thus, there is an equatorial trough, easterly jet at the surface on either side of the equator & the expected Hadley cell (Fig. 5 d). Note also in Fig. 5 that the planetary wave regime to the west of the forcing region is only one–third the size of the Kelvin wave regime because of the low wave speed (see Table 1).

When the heating is distributed asymmetrically about the equator [In this case $Q(x.y)>0$ on northern hemisphere and $Q(x.y)<0$ on Southern Hemisphere; thus, heating on north side and cooling on south side of the equator, see eq. (2.21)], the vertical perturbation velocity are found to be positive with centers located on both sides of the equator (Fig. 6 a). Meanwhile the low pressure center is on the Northern hemisphere and high pressure center on the Southern hemisphere, both correspond to heating region and cooling region, respectively (Fig. 6 b). The cross equatorial flow is found clearly from southern to Northern hemisphere (Fig. 6 a,b). Also, north of the equator, westerly is in the lower

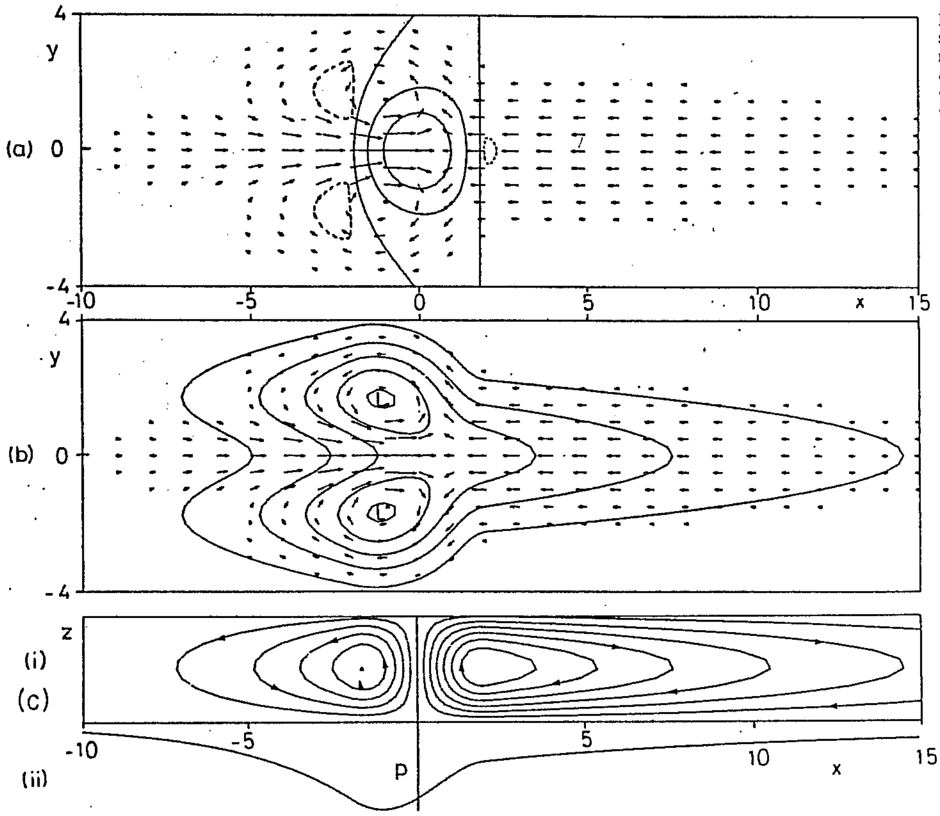
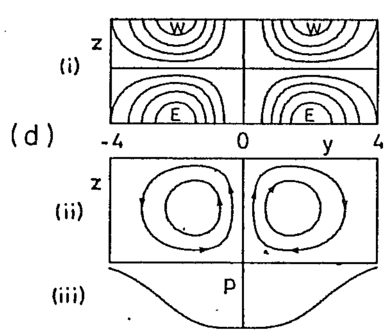


Figure 5 Solution for heating symmetric about the equator in the region $|x| < 2$ for decay factor $\varepsilon = 0{\cdot}1$.
(a) Contours of vertical velocity w (solid contours are 0, 0·3, 0·6, broken contour is −0·1) superimposed on the velocity field for the lower layer. The field is dominated by the upward motion in the heating region where it has approximately the same shape as the heating function. Elsewhere there is subsidence with the same pattern as the pressure field.

(b) Contours of perturbation pressure p (contour interval 0·3) which is everywhere negative. There is a trough at the equator in the easterly régime to the east of the forcing region. On the other hand, the pressure in the westerlies to the west of the forcing region, though depressed, is high relative to its value off the equator. Two cyclones are found on the north-west and south-west flanks of the forcing region.

(c) The meridionally integrated flow showing (i) stream function contours, and (ii) perturbation pressure. Note the rising motion in the heating region (where there is a trough) and subsidence elsewhere. The circulation in the right-hand (Walker) cell is five times that in each of the Hadley cells shown in (c).



(d) The zonally integrated solution showing (i) contours of zonal velocity (E marks the core of the easterly flow and W of the westerly flow aloft), (ii) stream function contours, and (iii) perturbation pressure.

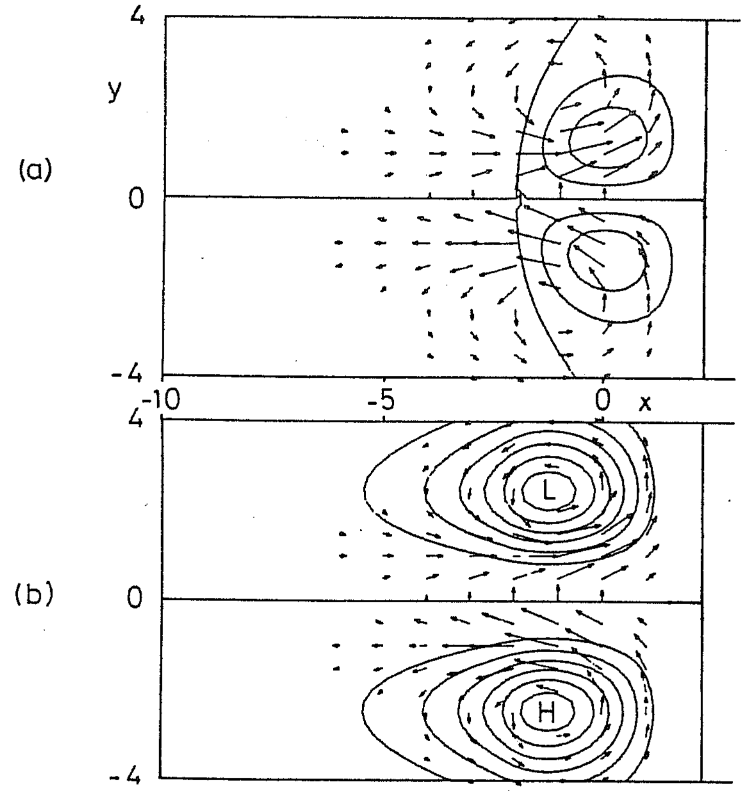
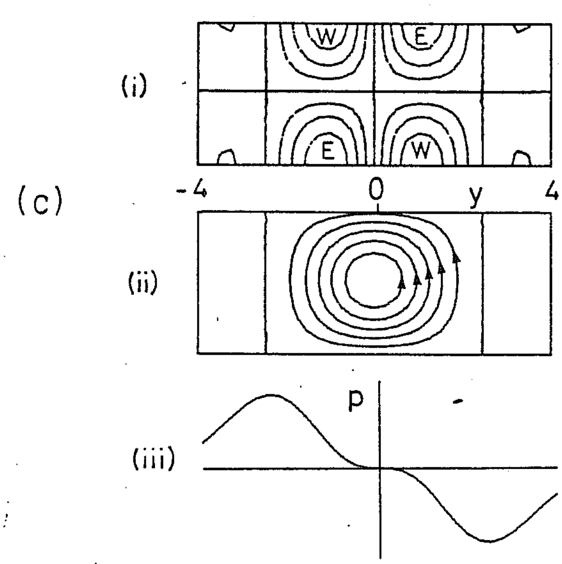


Figure 6 Solution for heating antisymmetric about the equator in the region $|x| < 2$ for decay factor $\varepsilon = 0{\cdot}1$.
(a) Contours of vertical velocity w (contour interval 0·3) superimposed on the velocity field for the lower layer. The field is dominated by the motion in the heating region where it is approximately the same shape as the heating function (positive in northern hemisphere). Outside the forcing region, the pattern is the same as for the pressure field with subsidence in the northern hemisphere and weak upward motion in the southern hemisphere. There is no motion for $x > 2$.
(b) Contours of perturbation pressure p (contour interval 0·3) which is positive in the north (where there is an anticyclone) and negative south of the equator (where there is a cyclone). The flow has the expected sense of rotation around the pressure centres and flows down the pressure gradient where it crosses the equator. All fields are zero for $x > 2$.



(c) The zonally integrated solution showing (i) contours of eastward velocity, (ii) stream function contours, and (iii) perturbation pressure.

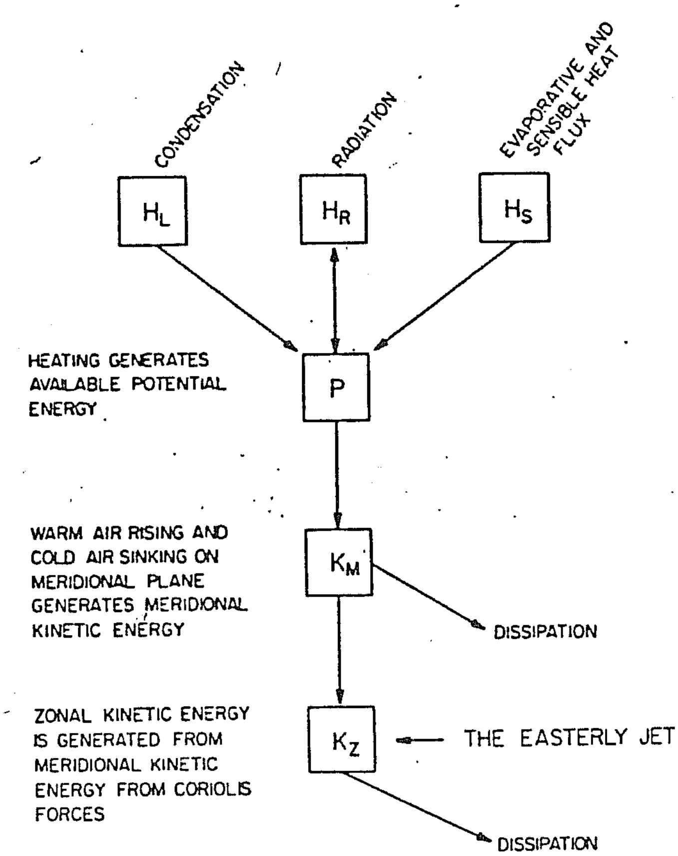


Figure 7 Energy exchanges in a zonally symmetric model simulation of the monsoon. The three main forms of heating [H_L, the condensation heating; H_R, the radiative heating (which is negative when it is cooling); and H_S, the sensible heating from the ocean] contribute to a net generation of available energy P. K_m denotes the meridional kinetic energy produced by vertical circulations, that is, by the ascent of warm air over land area and the descent of relatively colder air over the ocean to the south. The direction of the arrow from P to K_m represents the conversion from available-potential to kinetic energy. The exchange of energy from K_m to K_z (K_z denotes zonal kinetic energy, i.e., kinetic energy of west to east motions) occurs as air traveling northward or southward is deflected by the Coriolis forces. Finally, there is dissipation that depletes both zonal and meridional energy.

layer and easterly in the higher layer, south of the equator, they show the opposite case. A dominant Hadley cell in this case is found such that the rising branch is on the north and sinking branch on the south (Fig. 6 c). Note that this phenomena is consistent with the heating distribution on both side of the equator. However, one may notice that an upward motion center just happens to be over the cooling region on the south side of the equator, this is somehow inconsistent with the results analyzed above! Also notice that in the case of asymmetric heating, no motion is found east of the heating/cooling regions (Fig. 6). In both modes (n=0 and n=2). Solutions is zero for x>L (see Table 1). Therefore, the asymmetric heating scheme is somewhat questionable in the sense that tropical Pacific disturbance of the atmosphere is rarely found to be at rest.

Basically what is achieved in the antisymmetric case is a description of the following features:

- Mascarene High
- South east trades of the southern hemisphere
- Cross equatorial flow
- South west monsoon flow
- Monsoon trough
- Heating (i.e. rainfall) in the trough
- Rising motion in the trough
- Implied warm troposphere
- Tibetan (or upper) anticyclone
- Easterly flow south of the upper anticyclone

These were the basic ingredients of the summer monsoon that we had outlined in section 1. The same degree of description of the Winter monsoon system is also obtained from the symmetric heating.

5. ACKNOWLEDGEMENTS

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