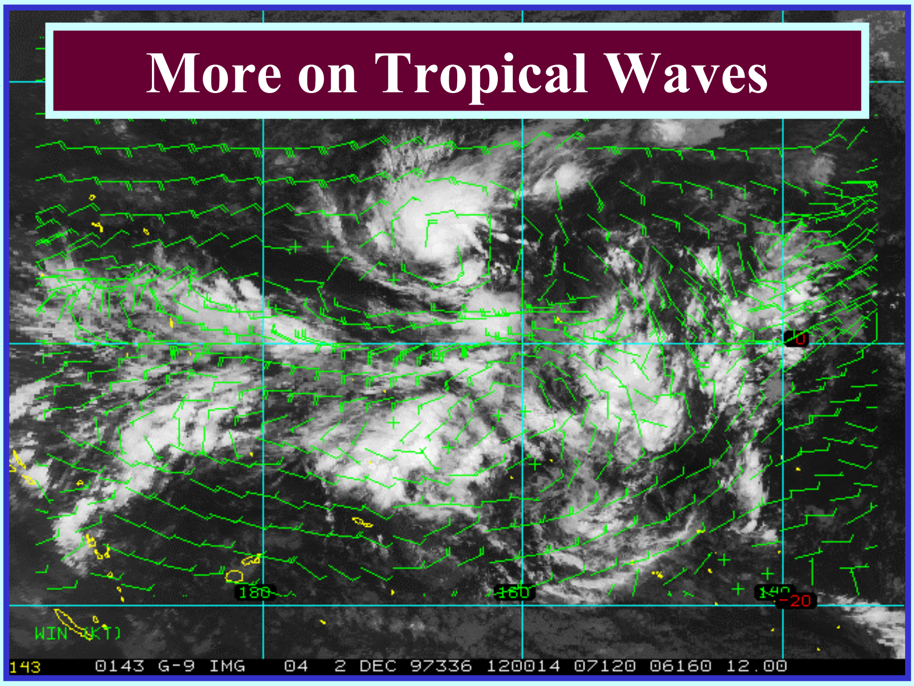
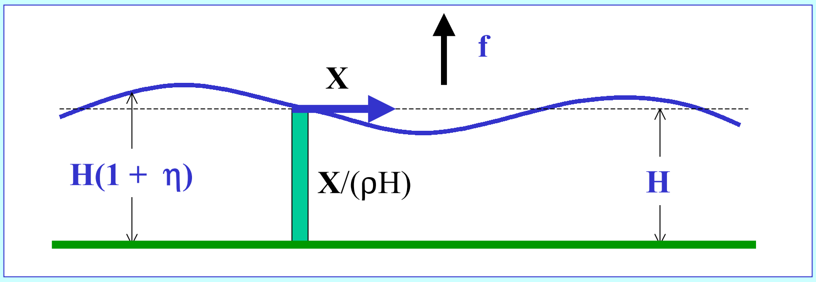
More on Tropical Waves



Topics

- Steady forced motions
- Large-scale response of the tropical atmosphere to transient convection
- A theory for midlatitude forcing of tropical motions during winter monsoons

Response to steady forcing



- Consider a homogeneous ocean layer of mean depth H forced by a surface wind stress **X** = (X,Y) per unit area.
- Assume that wind stress is distributed uniformly with depth as a body force **X**/(ρH) per unit mass.
- Suppose that there is also a drag per unit mass acting on the water, modelled by the linear friction law - ru per unit mass.
- See e.g. A. E. Gill (1982): Atmosphere – Ocean Dynamics.

Equations

Friction

$$-\beta y v = -gH\partial_x \eta + X/(\rho H) - ru$$

$$\beta y u = -gH\partial_y \eta + Y/(\rho H) - rv$$

$$c^2(\partial_x u + \partial_y v) = -gE/\rho - c^2 r\eta$$

Evaporation/heating

Eliminate u and η

$$\frac{r}{c^2}\left(r^2 + f^2\right)v - r\left(\frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial y^2}\right) - \beta\frac{\partial v}{\partial x} =$$

$$\frac{1}{\rho H}\left\{\frac{r}{c^2}\left(rY - fX\right) + r\frac{\partial E}{\partial y} - \frac{\partial}{\partial x}\left(\frac{\partial Y}{\partial x} - \frac{\partial X}{\partial x} + fE\right)\right\}$$

$$\frac{r}{c^2}\cancel{\left(r^2+f^2\right)}v - r\cancel{\left(\frac{\partial^2 v}{\partial x^2}+\frac{\partial^2 v}{\partial y^2}\right)} - \beta\frac{\partial v}{\partial x} =$$

$$\frac{1}{\rho H}\left\{\cancel{\frac{r}{c^2}\left(rY - fX\right)} + \cancel{r\frac{\partial E}{\partial y}} - \frac{\partial}{\partial x}\left(\frac{\partial Y}{\partial x} - \frac{\partial X}{\partial x} + fE\right)\right\}$$

Case of small friction: $r \to 0$

$$\beta\frac{\partial v}{\partial x} = \frac{1}{\rho H}\frac{\partial}{\partial x}\left(\frac{\partial Y}{\partial x} - \frac{\partial X}{\partial x} + fE\right)$$

Integrate w. r. t. x

$$\beta v = \frac{1}{\rho H}\left(\frac{\partial Y}{\partial x} - \frac{\partial X}{\partial x} + fE\right)$$

$\Leftarrow$ **when** $E = 0 \Rightarrow$
Sverdrup's formula

$$\frac{r}{c^2}\left(r^2+f^2\right)v - r\left(\cancel{\frac{\partial^2 v}{\partial x^2}}+\frac{\partial^2 v}{\partial y^2}\right) - \cancel{\beta\frac{\partial v}{\partial x}} =$$

$$\frac{1}{\rho H}\left\{\frac{r}{c^2}\left(rY - fX\right) + r\frac{\partial E}{\partial y} - \cancel{\frac{\partial}{\partial x}\left(\frac{\partial Y}{\partial x} - \frac{\partial X}{\partial x} + fE\right)}\right\}$$

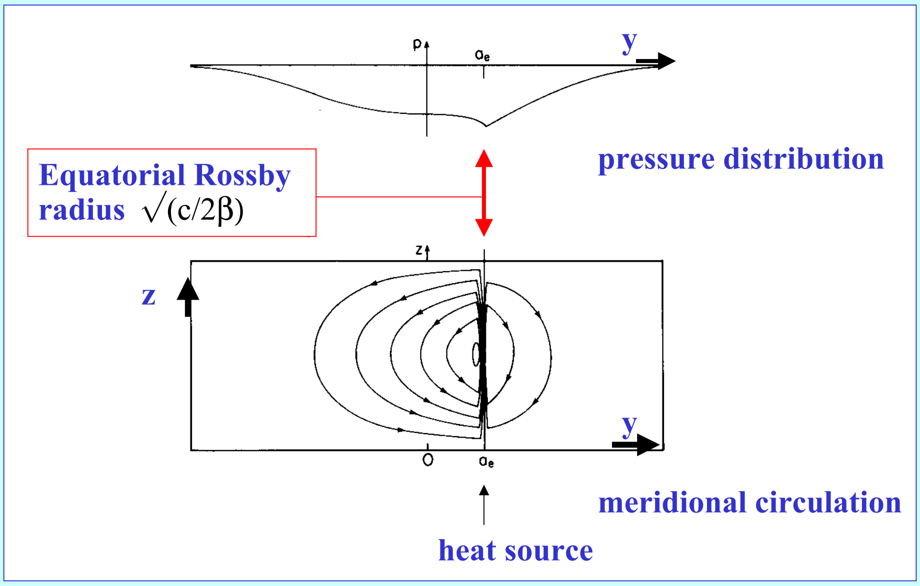
Zonally-independent flows: $\partial/\partial x = 0$

$$\frac{\left(r^2+f^2\right)}{c^2}v - \frac{\partial^2 v}{\partial y^2} - = \frac{1}{\rho H}\left\{\frac{\left(rY - fX\right)}{c^2} + \frac{\partial E}{\partial y}\right\}$$

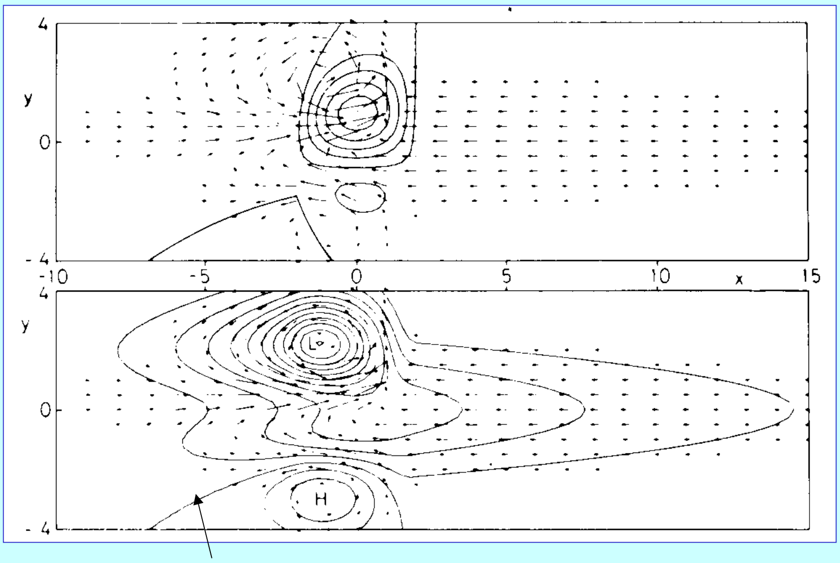
Valid on an f-plane or when $f = \beta y$

When $f = \beta y$**, solutions are in terms of parabolic cylinder functions of order** $^1/_2$ **(see Gill, 1982, p467).**

The linear solution due to a line source of heat/evaporation.



Heating confined to the range of longitudes $|x| < 2L_R$

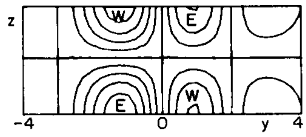


vertical velocity

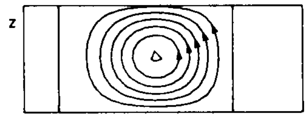
arrows = wind

Heating confined to the range of longitudes $|x| < 2L_R$

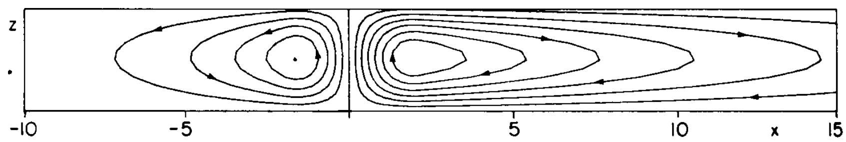
Zonal flow ⇒



Meridional flow ⇒



Meridionally-averaged Walker circulation

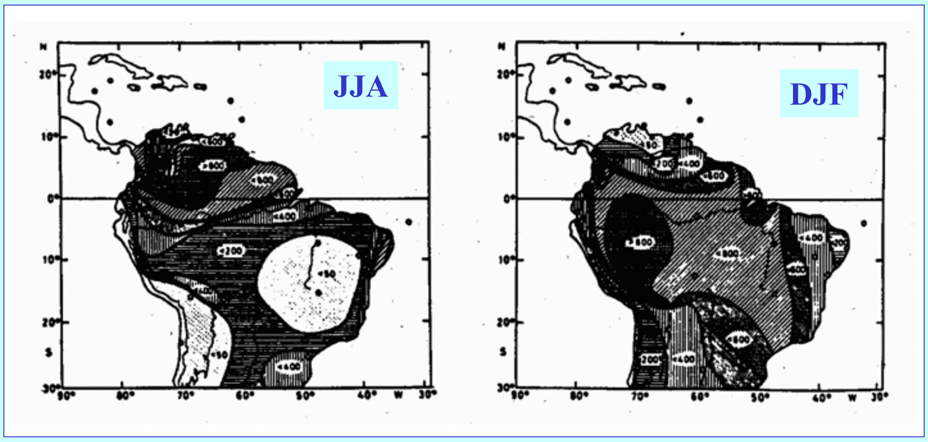


Large-scale response of the tropical atmosphere to transient convection

Silva Dias, Schubert & DeMaria, 1983

- **Large west-east asymmetries in the tropical circulations are associated with the equatorial continental areas of South America, Africa and the Maritime continent.**
- **SDSD show average 200 and 700 mb flow over South America for four winters (JJA) and four summers (DJF) between 1996 and 1970.**
- **A prominent feature is the Bolivian high above 200 mb.**
- **May be maintained by latent heat release.**

Rainfall in mm

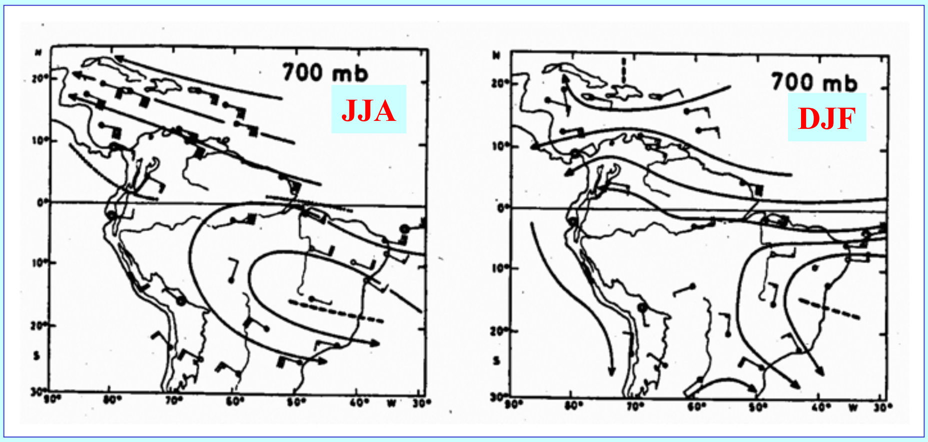


Winter

Summer

Silva Dias et al., Fig. 1

700 mb streamlines

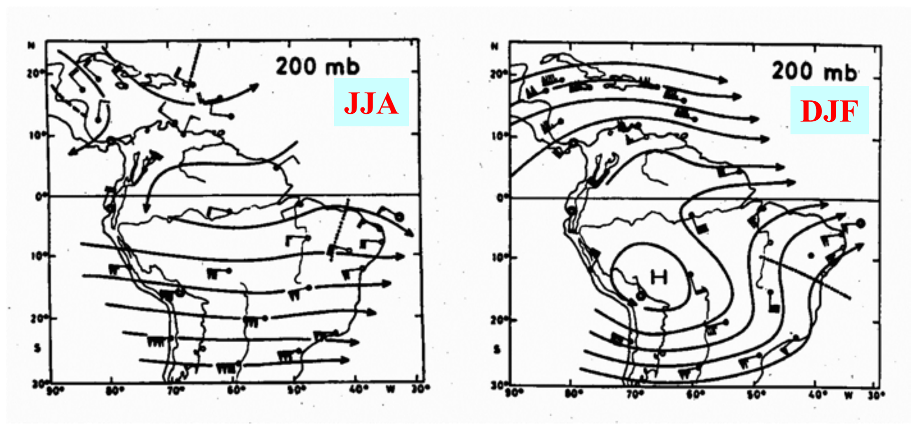


Winter

Summer

Silva Dias et al., Fig. 1

200 mb streamlines



Winter | Summer

Silva Dias et al., Fig. 1

(Linearized) Equations of motion

$$z = \ln(p_0 / p)$$

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - \beta y v = -\frac{\partial \phi}{\partial x} + F$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} + \beta y u = -\frac{\partial \phi}{\partial y} + G$$

Momentum sources

$$\frac{\partial \phi}{\partial z} - RT = 0$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} - w = 0$$

Heat source

$$\frac{\partial T}{\partial t} + u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y} + w\left(\frac{\partial T}{\partial z} + \kappa T\right) = \frac{\dot{Q}}{c_p}$$

$$\frac{\partial u}{\partial t} - \beta y v + \frac{\partial \phi}{\partial x} = \frac{\partial \tilde{u}}{\partial t}$$

$$\frac{\partial v}{\partial t} + \beta y u + \frac{\partial \phi}{\partial y} = \frac{\partial \tilde{v}}{\partial t}$$

$$\frac{\partial}{\partial t}\left[e^{z} \frac{\partial}{\partial z}\left(\frac{e^{-z}}{R\Gamma}\right)\frac{\partial \phi}{\partial z}\right] - \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = \frac{\partial}{\partial t}\left[e^{z} \frac{\partial}{\partial z}\left(\frac{e^{-z}}{R\Gamma}\right)\frac{\partial \tilde{\phi}}{\partial z}\right]$$

Basic state static stability: $\Gamma(z) = \frac{d\overline{T}}{dz} + \kappa \overline{T}$

Forcing terms: $F = \frac{\partial \tilde{u}}{\partial t}$, $G = \frac{\partial \tilde{v}}{\partial t}$, $\frac{RQ}{c_p} = \frac{\partial}{\partial t}\left(\frac{\partial \tilde{\phi}}{\partial z}\right)$

Stürm-Liouville Transform: $f(x, y, z, t) = u, v, \phi, u_n, v_n, \phi_n$

$$f_n(x, y, t) = \int_0^{z_T} f(x, y, z)\Psi_n(z)e^{-z/2}dz$$

Inverse: $f(x, y, z, t) = \sum_{n=0}^{\infty} f_n(x, y, z)\Psi_n(z)e^{z/2}$

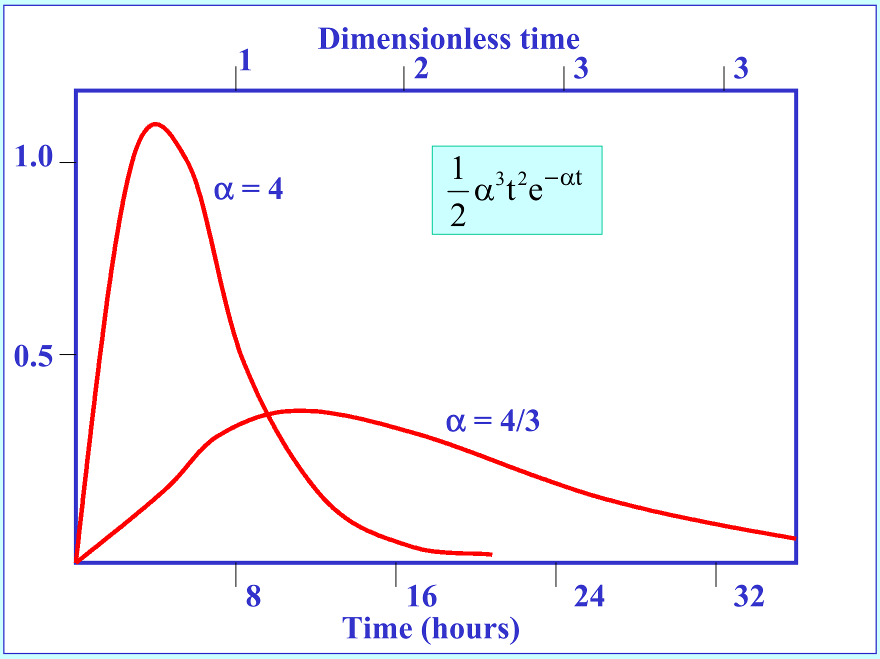
Transformed equations

$$\frac{\partial u_n}{\partial t} - \beta y v_n + \frac{\partial \phi_n}{\partial x} = \frac{\partial \tilde{u}_n}{\partial t}$$

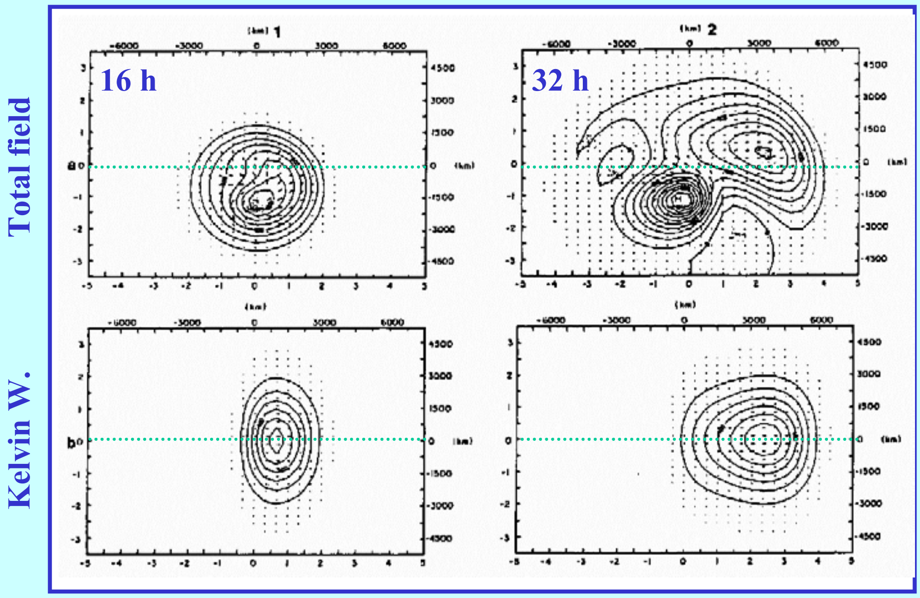
$$\frac{\partial v_n}{\partial t} + \beta y u_n + \frac{\partial \phi_n}{\partial y} = \frac{\partial \tilde{v}_n}{\partial t}$$

$$\frac{\partial \phi_n}{\partial t} + c_n^2\left(\frac{\partial u_n}{\partial x} + \frac{\partial v_n}{\partial y}\right) = \frac{\partial \tilde{\phi}_n}{\partial t}$$

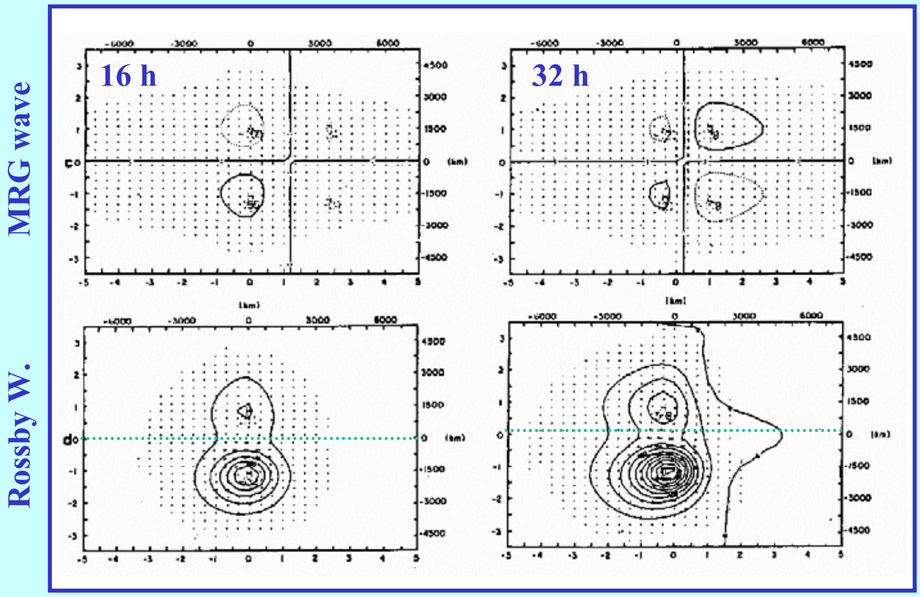
$c_n = (gh_n)^{1/2}$



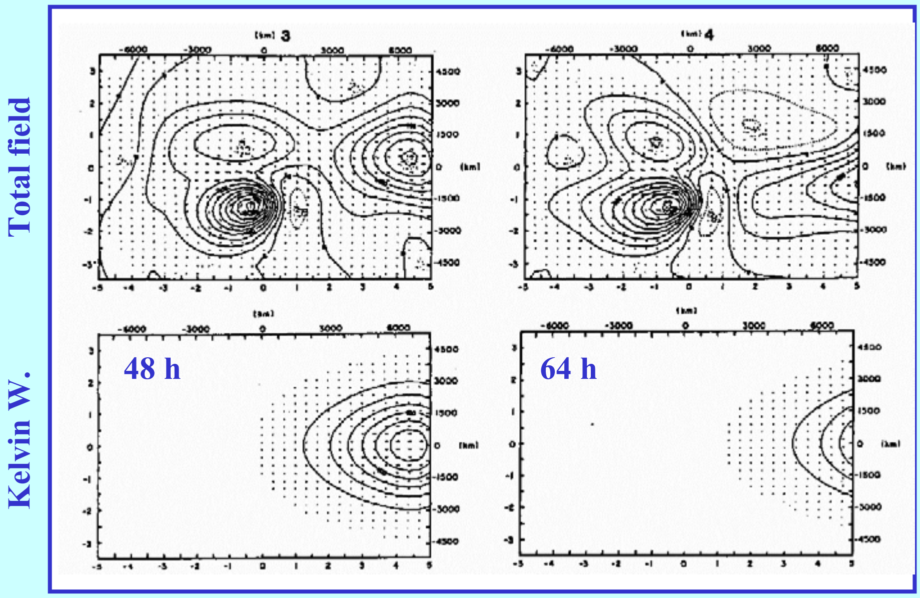
Wind and geopotential



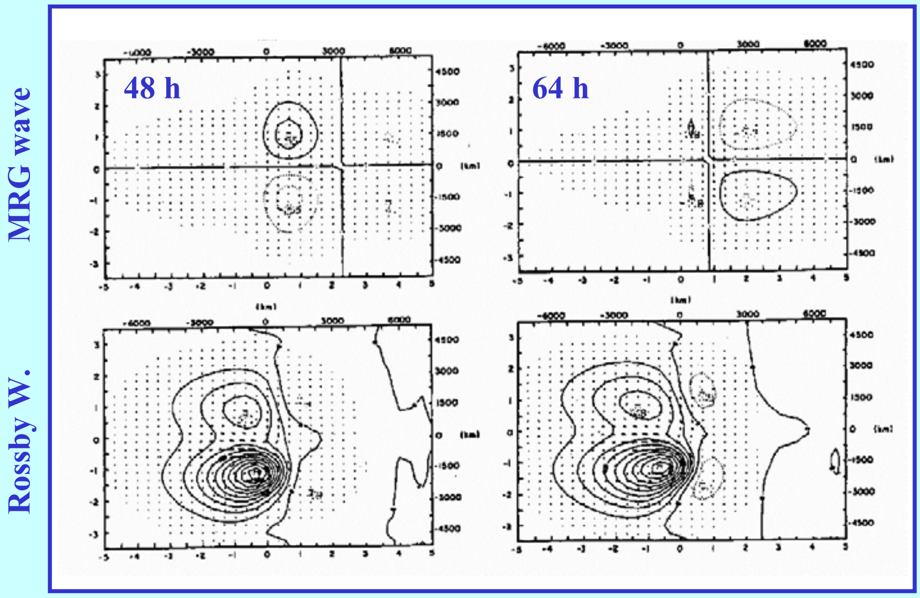
Wind and geopotential



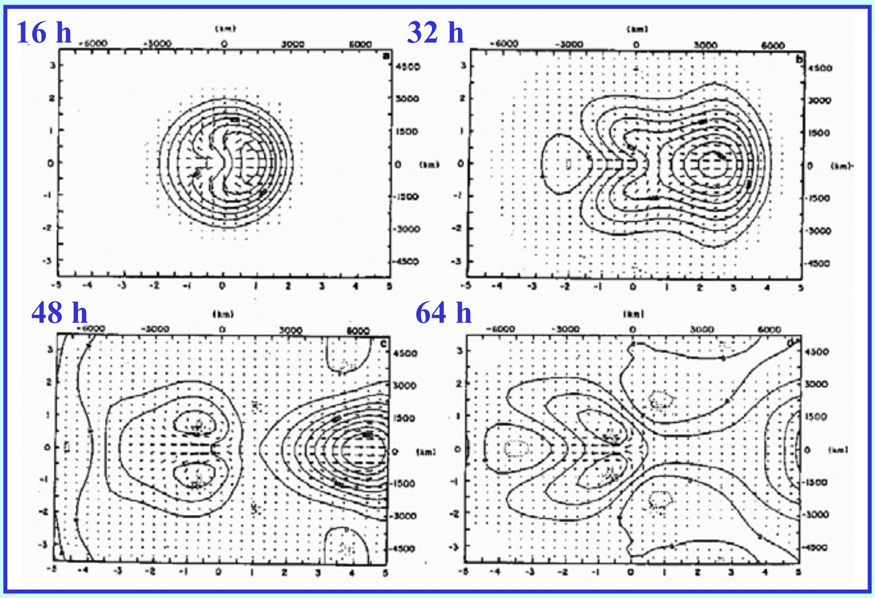
Wind and geopotential



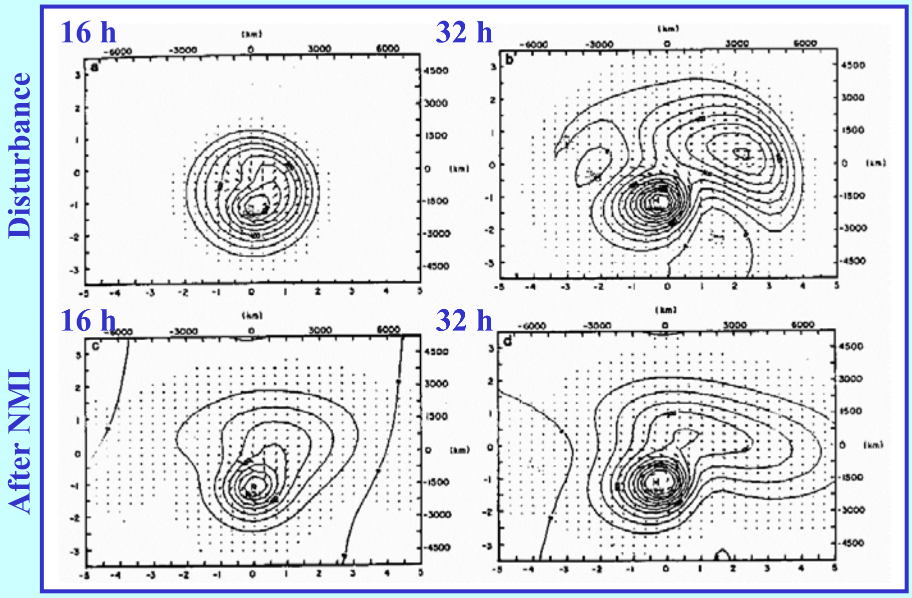
Wind and geopotential



Wind and geopotential



Wind and geopotential



Summary - 1

- The amount of Kelvin wave energy generated is greater when the forcing is rapid and when the heat source is close to the equator.
- The convection that occurs over the South American continent and also over the Indonesian region has large variations on short time scales.
- Since both of these regions are close to the equator, there should be a large Kelvin wave response to this convective forcing.

Summary - 2

- At upper levels, convergence occurs on the eastern side of the Kelvin wave group with divergence on the western side.
- This convergence-divergence pattern and associated vertical motion field may be important for modulating convection in areas far to the east of the forcing.
- There is some observational evidence for this modulation of convection by the Kelvin waves.

Summary - 3

- The generation of Kelvin waves by transient convection has some implications for the initialization of tropical forecast models.
- The basic idea of many nonlinear normal mode initialization procedures is to obtain the slow mode part of an initial field from observations, and then to diagnose the fast mode part in such a way that the initial tendency of the fast modes is zero.
- When this type of procedure is applied, the magnitude of die nonlinear interaction of the slow modes and the magnitude of the forcing terms at the initialization time determine the amplitude of the fast modes.

Summary - 4

- For the case of transient convection, the amplitude of the Kelvin modes will be fairly large after the amplitude of the forcing has decreased.
- Since the Kelvin waves are considered to be fast modes (except for very long waves), it may not be possible to diagnose these after the magnitude of the forcing has decreased and the waves have propagated away from the source region.

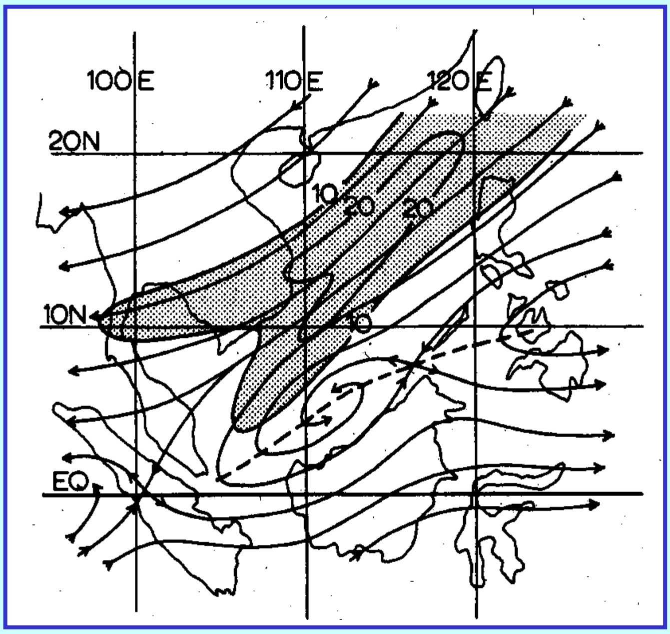
A theory for midlatitude forcing of tropical motions during winter monsoons

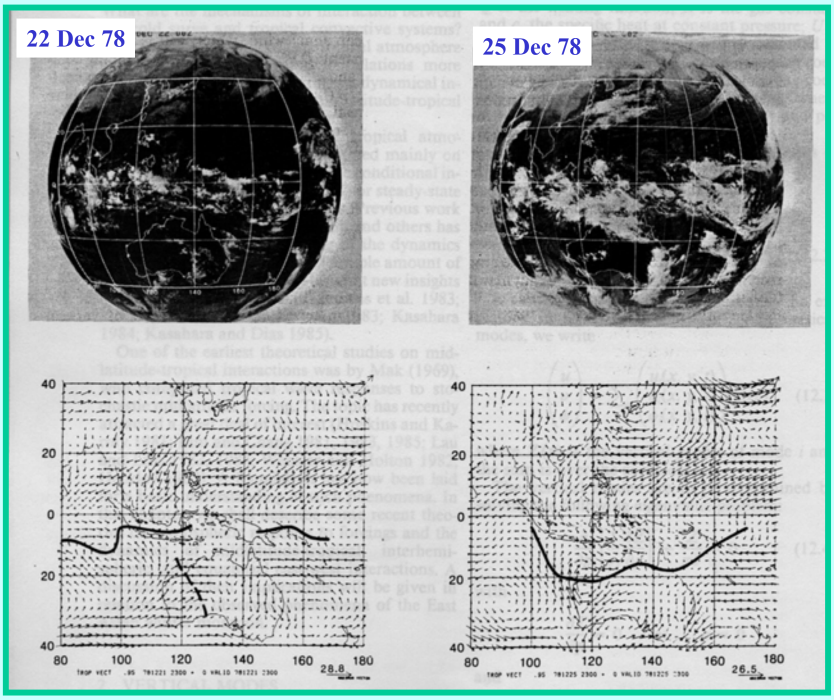
Lim. H, C.-P. Chang, 1981

- In the mid1atitudes, a surge in the northeast trades arrives with a steep rise of surface pressure, a sharp drop of temperature and a strengthening of northerly winds.
- The cold front leading the surge sometimes brings stratus and rain but a strong surge is generally associated with subsiding motions which leads to clearing of weather (Danielsen and Ho, 19692; Ramage, 1971).
- Although the front associated with a surge cannot normally be followed southward of about 25^0 N.

- The surge propagates equatorwards in dramatic fashion, ss a vigorous surge reaches the South China coast, northerly winds freshen almost simultaneously several hundred kilometres to the south, far beyond the region where the winds could have pushed the front. (Ramage, 1971; Chang *et al*., 1979).
- A belt of strong northeasterly winds from within 24 h off the South China and Vietnam coasts.

Northeasterly surges





Wind surges

Cross-Equatorial Influence of Winter Hemisphere Subtropical Cold Surges

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(Manuscript received 18 April 1983, in final form 9 April 1985)

ABSTRACT

Two case studies are presented of winter hemisphere cold surges affecting the summer hemisphere tropics. One case has been chosen from each of the Northern and Southern Hemispheres. In both instances, the subtropical rise in pressure is tracked equatorward. This leads to a pressure rise at the equator and the establishment of a west–east pressure gradient at low latitudes in the opposite hemisphere. These effects lead in turn to an enhanced cross-equatorial component of flow and enhanced monsoon westerly flow in the summer hemisphere. Both enhanced wind flow effects are observed through a deep layer from the surface up to at least 500 mb. In both cases, the sequence of events also includes the development of a tropical cyclone in the summer hemisphere monsoon trough.

The generality of the above sequence of events is investigated with time series data for several seasons. It is shown that day-to-day changes in equatorial pressure are significantly correlated to pressure changes in the winter hemisphere subtropics. It is also shown that the strength of the low-latitude westerly winds is well correlated with the synoptic scale west–east pressure difference near the equator.

A shallow water model calculation for surges

Based on:

On the dynamics of midlatitude-tropical interactions and thew winter monsoon, by Hoch Lim and C.-P. Chang.

In Monsoon Meteorology, Ed. C.-P. Chang and T. N. Krishnamurti.

Linearized equations on an equatorial β-plane

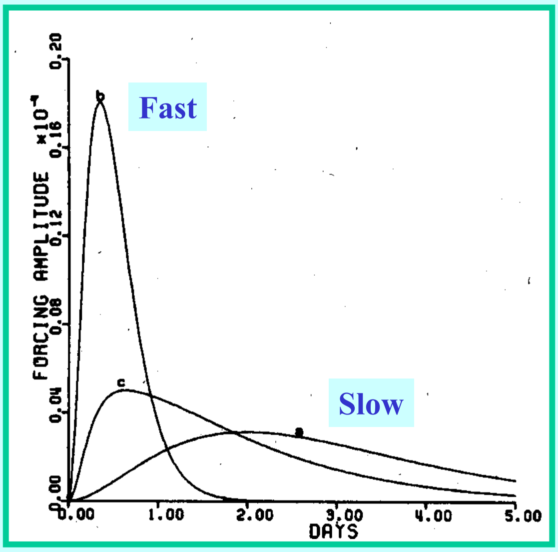
$$\frac{\partial u}{\partial t} - \beta y v + g\frac{\partial h}{\partial x} = 0$$

$$\frac{\partial v}{\partial t} + \beta y u + g\frac{\partial h}{\partial y} = 0$$

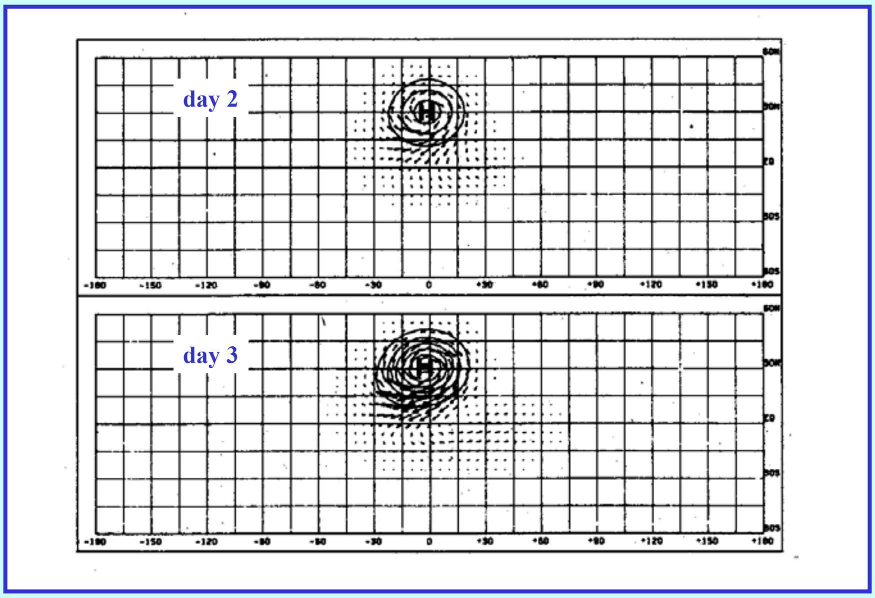
$$\frac{\partial h}{\partial t} + H\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = \Phi$$

Motions forced by a mass source term

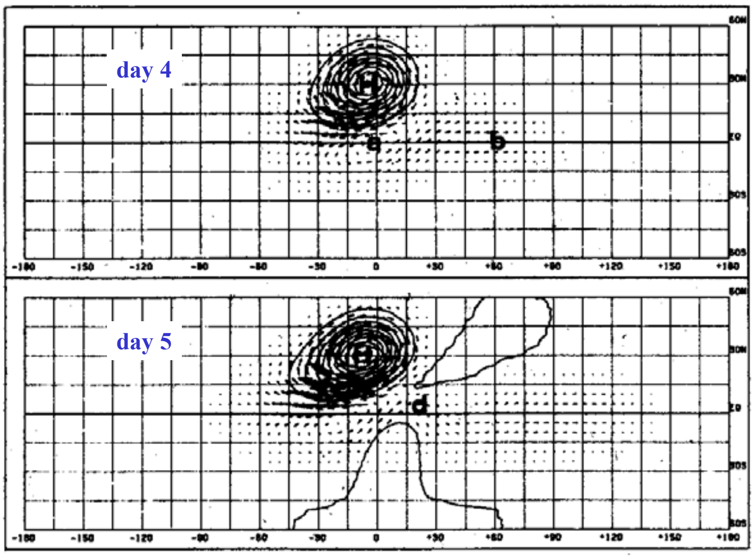
Variation of forcing amplitude with time



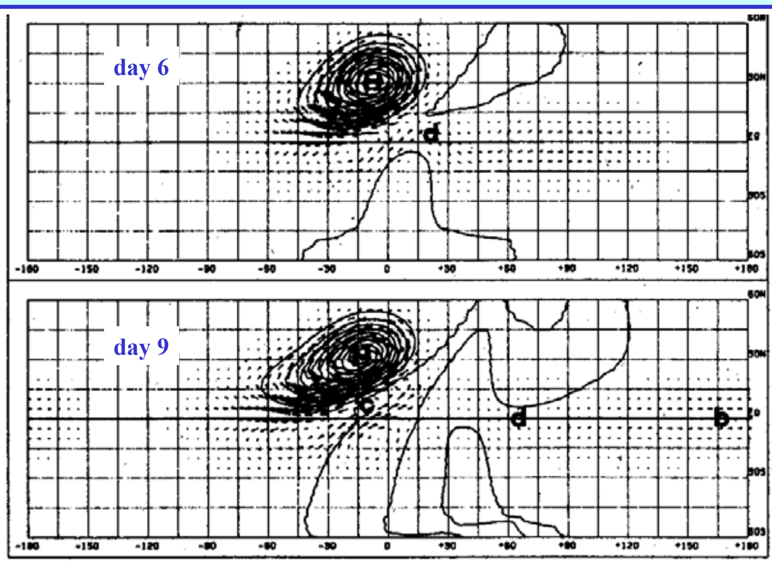
Slow forcing



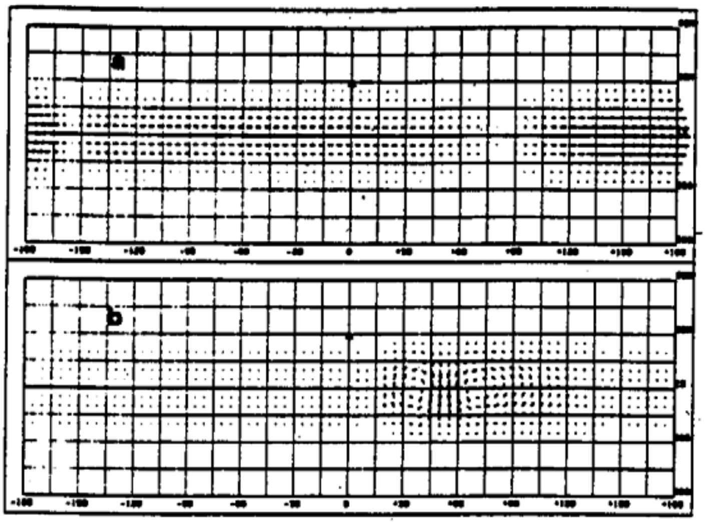
Slow forcing



Slow forcing

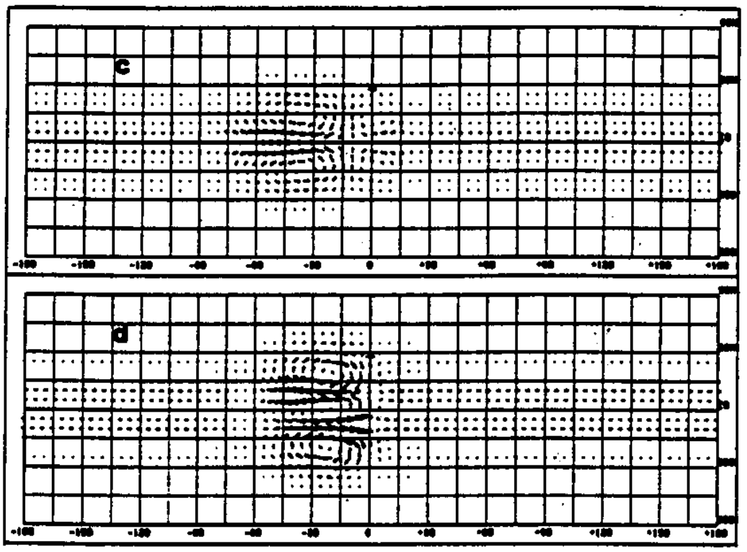


Kelvin wave group (n = −1)

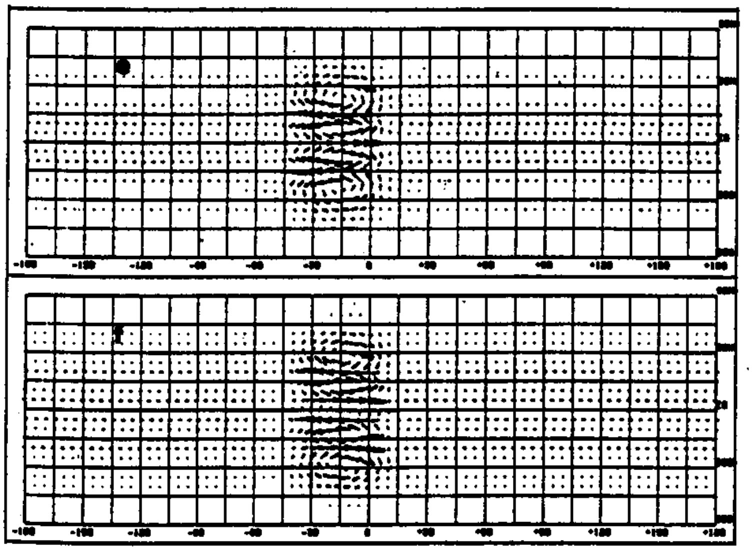


Mixed Rossby-gravity wave group (n = 0)

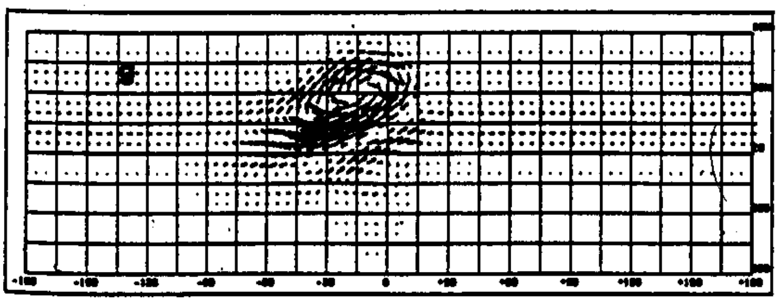
Rossby-gravity wave group (n = 1 – 2)

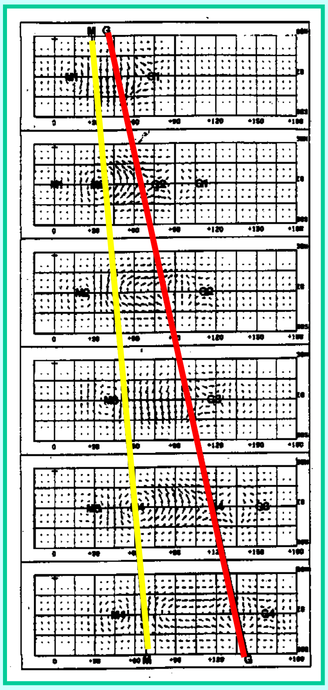


Rossby-gravity wave group (n = 3 – 4)



Combined Rossby-gravity wave group



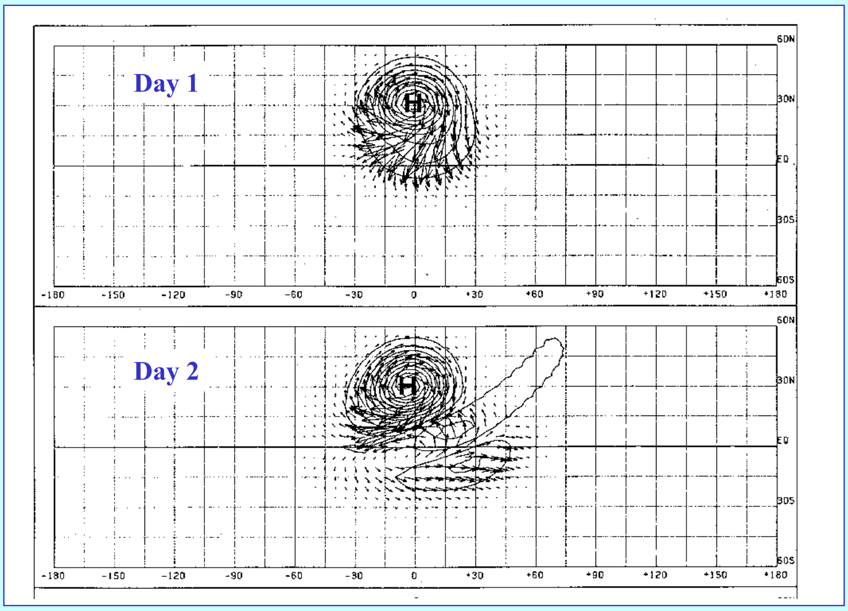


n = 0 Hermite modes excited by the fast forcing from day 4 to day 9

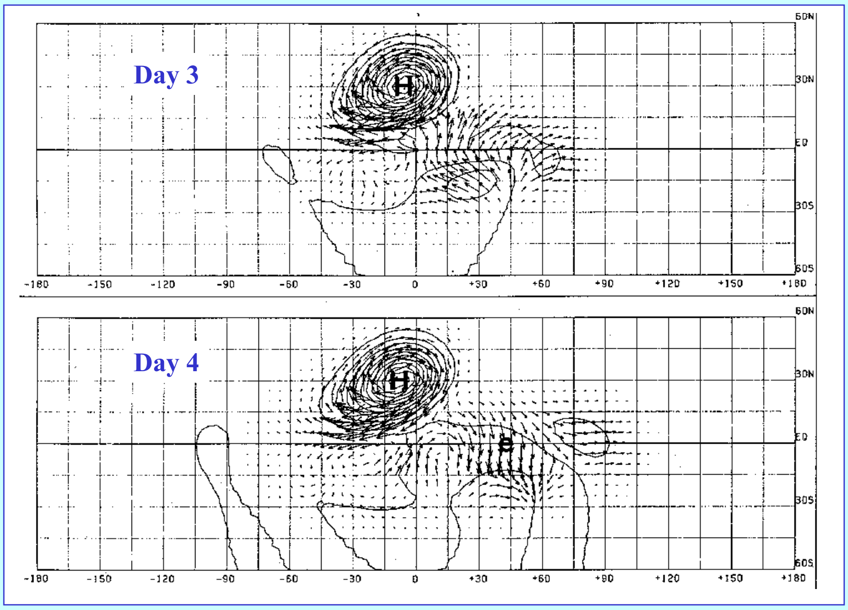
Trajectories of the centre (area of maximum amplitude) of the mixed Rossby-gravity wave group

Trajectories of the centre (area of maximum amplitude) of the inertia-gravity wave group

Fast forcing

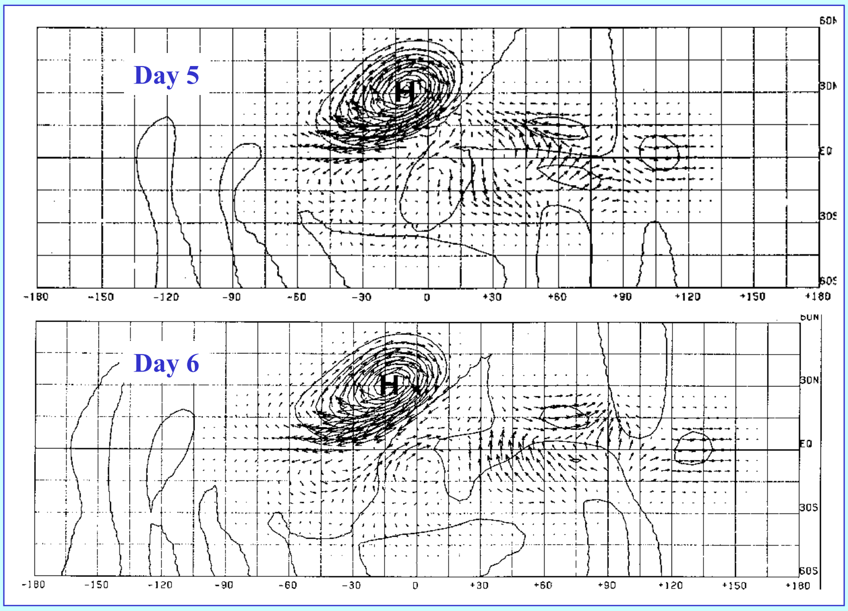


Day 3 - 4



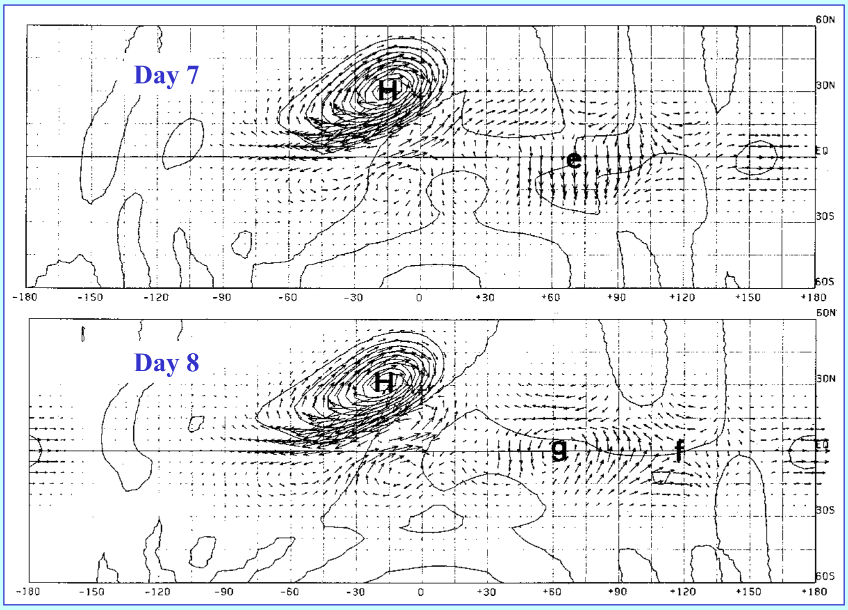
Day 3
H
Day 4
H
60N
30N
EQ
30S
60S
-180 -150 -120 -90 -60 -30 0 +30 +60 +90 +120 +150 +180

Day 5 - 6



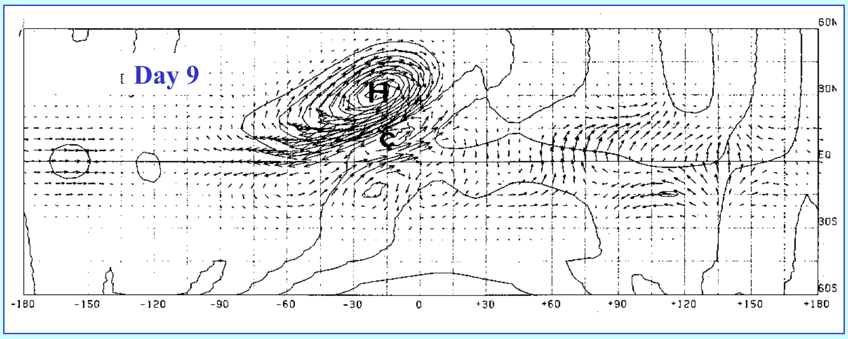
Day 5
H
Day 6
H
30N
EQ
30S
60S
-180 -150 -120 -90 -60 -30 0 +30 +60 +90 +120 +150 +180

Day 7 - 8



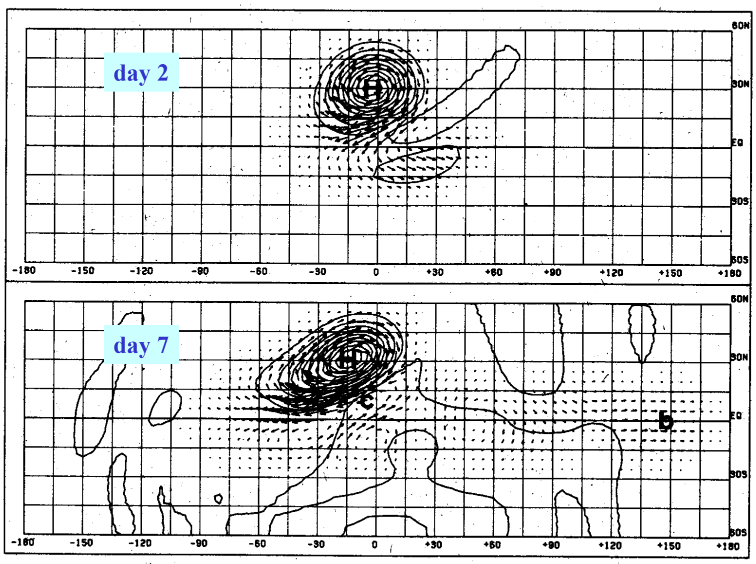
Day 7
H
e
Day 8
H
g
f
60N
30N
EQ
30S
60S
-180
-150
-120
-90
-60
-30
0
+30
+60
+90
+120
+150
+180

Day 9



Day 9
H
C
60N
30N
EQ
30S
60S
-180
-150
-120
-90
-60
-30
0
+30
+60
+90
+120
+150
+180

Realistic forcing



Summary - 1

- Midlatitude forcing with large time and space scales excite mainly Rossby and, to a much smaller amplitude, Kelvin waves.
- As either of these scales decreases, the amplitudes of mixed Rossby-gravity and inertia-gravity waves increase.
- Immediately following a pressure forcing, the strong wind surge is more northerly with a gravity wave character.
- The strong cross-isobaric divergent flow occurs without the presence of surface friction.
- It becomes more easterly as the Rossby wave group is established, in agreement with observations.

Summary - 2

- The development of the main tropical flow pattern after a midlatitude pressure forcing, including the strong northeast wind belt southeast of the main anticyclone and a cyclonic shear zone further southeast of it, is represented by the slowly moving Rossby wave group response.
- This Rossby wave group develops a pronounced northeast-southwest tilt which is caused by the differential westward movement of wave modes, with the lower modes having relatively faster group velocities.
- The equatorial cyclonic shear flow southeast of the main anticyclone is an inherent property of the Rossby wave group response even in the absence of orography.

Summary - 3

- It is consistent with the mean winter condition of a northeast-southwest equatorial trough along Borneo-Philippines, and with an enhanced cyclonic circulation over Borneo following surges.
- The surge forcing also gives rise to eastward moving wave groups of the Kelvin, mixed Rossby-gravity, and inertia-gravity (mainly $n = 0$) modes.
- This result suggests a possible interpretation for the eastward moving cloud patterns observed recently during winter.

Summary - 4

- The Kelvin waves forced by midlatitude forcing. as is the case for the thermally forced Kelvin waves, have more energy in the larger wavelengths for a white noise random forcing.
- It appears that the diverse events of the monsoon surge might be given a coherent interpretation in terms of simple, inviscid, barotropic β-plane dynamics.
- However, many important questions remain …!

References

- Lim, H., C.-P. Chang, 1981: A theory for midlatitude forcing of tropical motions during winter monsoons. *J. Atmos. Sci.*, 38, 2377-2392 .
- Lim, H., C.-P. Chang, 1983: Dynamics of teleconnections and Walker circulations forced by equatorial heating. *J. Atmos. Sci.*, 38, 2377-2392 .
- Silva Dias, P. L., W. H. Schubert and M. DeMaria, 1983: Large-scale response of the tropical atmosphere to transient convection. *J. Atmos. Sci.*, 40, 2689-2707.

Dynamical Aspects of Twin Tropical Cyclones Associated with the Madden–Julian Oscillation

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(Manuscript received 22 July 1994, in final form 25 September 1995)

ABSTRACT

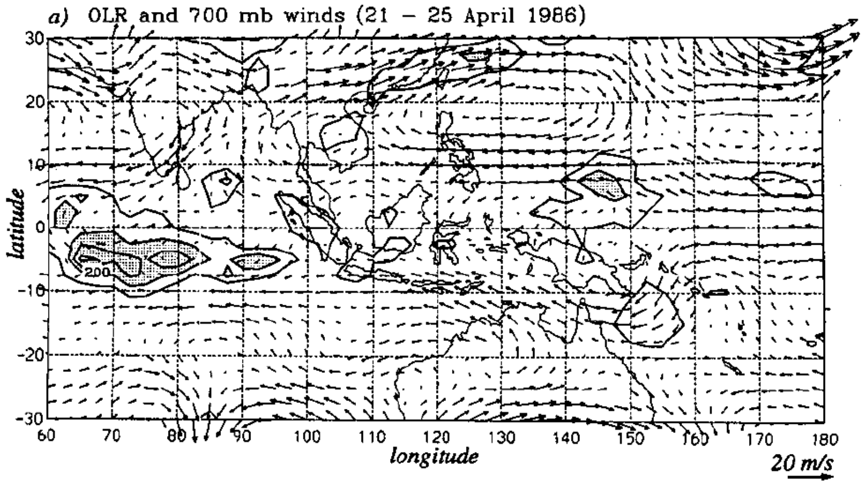
A nonlinear shallow-water model on the sphere is used to study barotropic aspects of the formation of twin tropical disturbances by Madden–Julian oscillation (MJO) convection.

In the model, the effect of MJO convection upon the lower-tropospheric tropical circulation was simulated by an eastward moving, meridionally elongated mass sink straddling the equator. The intensity and propagation speed of the mass sink were chosen to simulate observations that MJO convection intensifies while nearly stationary in the eastern equatorial Indian Ocean, weakens while moving eastward over the Maritime Continent, again intensifies once it reaches the west Pacific Ocean, and finally becomes stationary and dies off near the date line. This mass sink produced twin cyclones in the two regions where it was stationary, namely, where it was initially turned on and where it was turned off. In addition, the mass sink produced two zonally elongated cyclonic potential vorticity anomalies straddling the equator in the region where it propagated eastward.

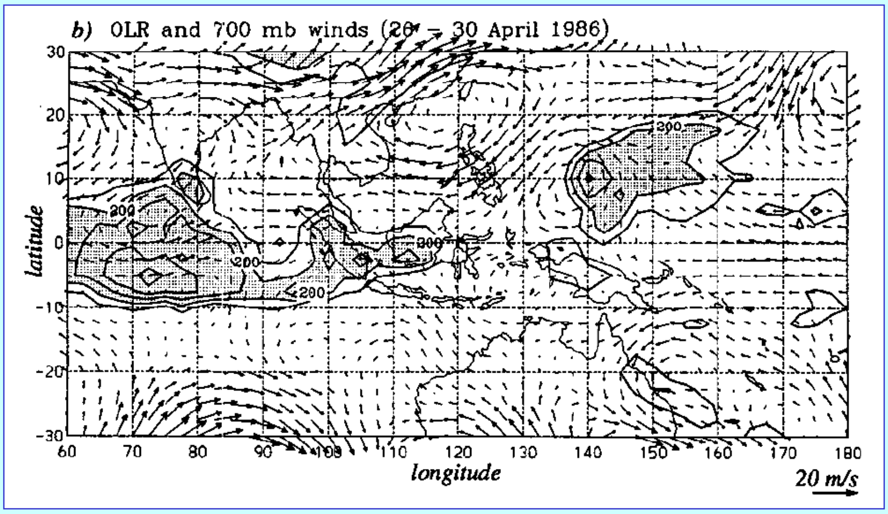
It is proposed that MJO convection produces twin tropical disturbances in the two regions where it is nearly stationary, namely, its region of formation in the eastern Indian Ocean and its region of decay near the date line. Additional tropical disturbances may arise from the breakdown of the elongated shear regions produced by the eastward propagating MJO convection.

In addition, a series of initial value experiments was performed to determine the conditions under which twin cyclones become so strongly coupled that they propagate directly eastward as a cyclone pair. Apparently, such movement requires the cyclones to be so close together that the situation rarely, if ever, occurs in nature.

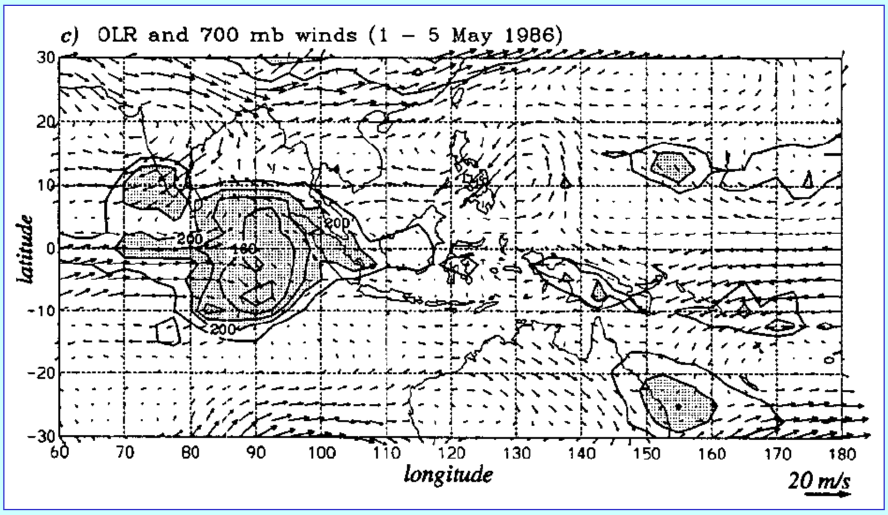
OLR and 700 mb winds 21 – 25 April 1986



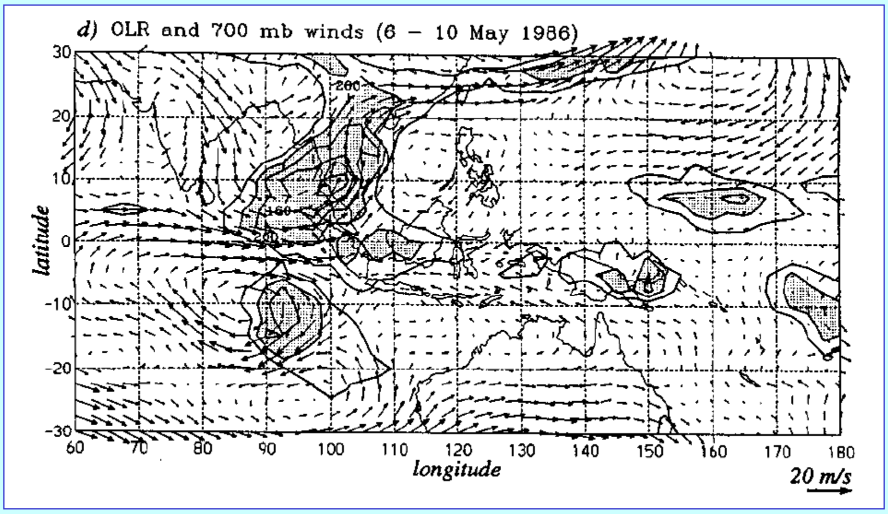
OLR and 700 mb winds 26 – 30 April 1986



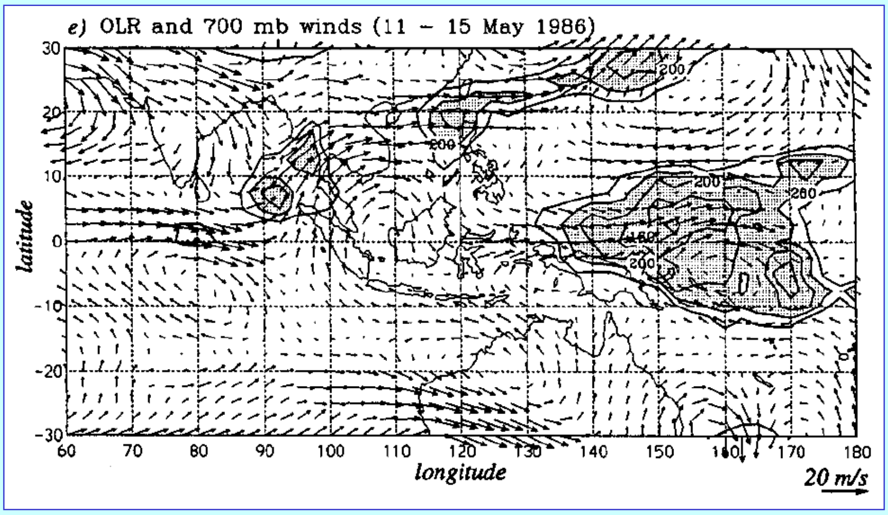
OLR and 700 mb winds 01 – 05 May 1986



OLR and 700 mb winds 06 – 10 May 1986



OLR and 700 mb winds 11 – 15 May 1986



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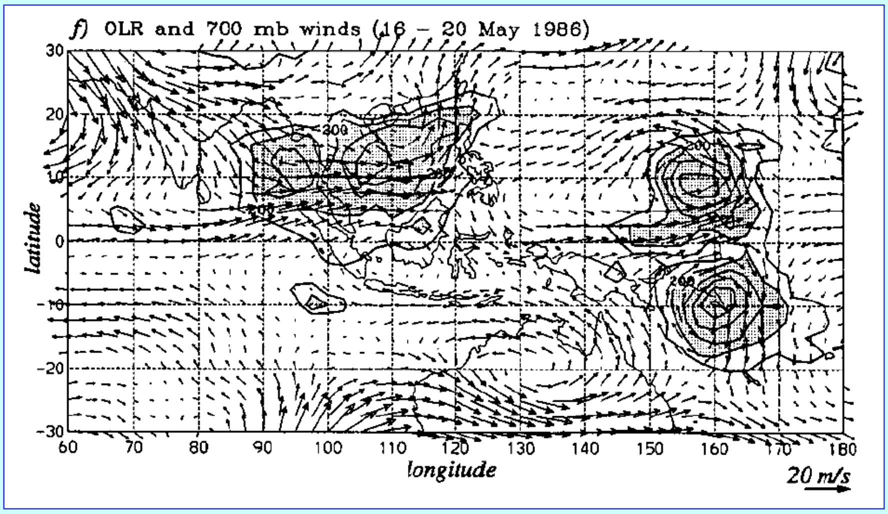
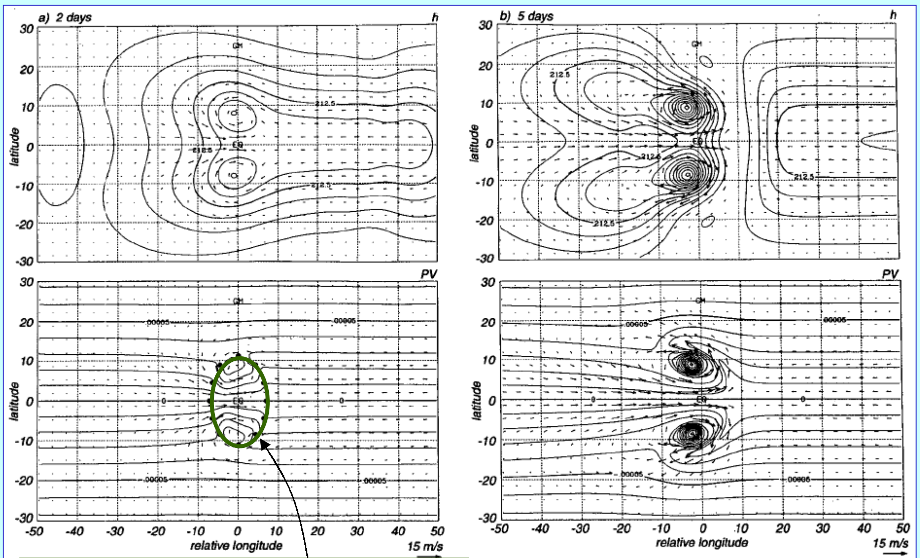
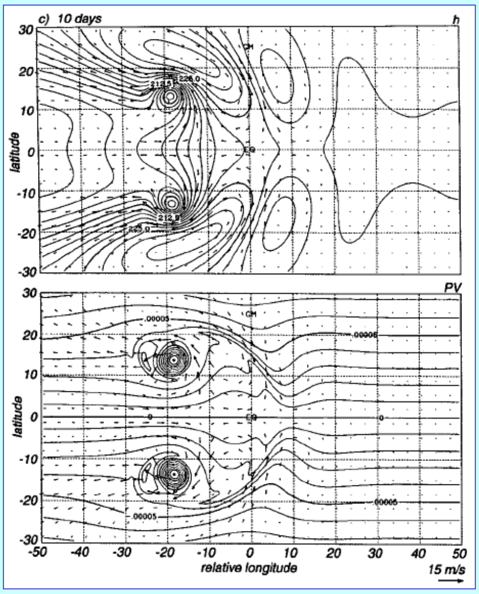


A shallow water model calculation



Stationary mass sink = convective heating

A shallow water model calculation



Summary

1. These calculations indicate the complexity of the response of the tropics to forcing.
2. The response is in the form of various wave types that can exist on the equatorial wave-guide.
3. The interaction of these waves with convection is still not well understood.

A relevant paper

A review of 'synoptic to intraseasonal' tropical waves with relevance to forecasting

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http://www.meteo.physik.uni-muenchen.de/~roger/Tropical_Meteorology/MatWheeler2001.pdf

The End