Monsoons in a Moist Axially Symmetric Model of the Atmosphere

Giulio Boccaletti

1 Introduction

Large scale meridional overturning circulations such as the Asian monsoon and the Hadley Cell are a dominant feature of the dynamics of the tropical atmosphere.

Our theoretical understanding of these circulations is largely based on idealized axisymmetric models, which describe them as steady-state and quasi-inviscid. In such theoretical models it can be shown [1] that meridional flow can only occur if it conserves angular momentum, entailing a condition on the thermal forcing of the atmosphere.

In the case of the Asian monsoon, Xie and Saiki [2] have recently argued that this constraint can be only met once moisture has penetrated over the Indian sub-continent, and that this is the result of a zonally asymmetric instability that draws water over land from the Indian Ocean, marking the onset of the monsoon.

Their argument entails that differences in vertical thermal structure between a moist convecting atmosphere and a dry one are critical for the existence of a monsoon. Furthermore, an axisymmetric model should be unable to produce a monsoon, as zonally asymmetric disturbances are absent.

To verify whether this is indeed the case, we construct an axisymmetric model of the atmosphere with interactive moist convection. The objective is to achieve a reasonable representation of the effects of moist convection on the large scale dynamics of the tropical atmosphere, therefore improving previous idealized models, in which typically a vertical thermal structure is externally imposed [3, 1].

Section (2) is devoted to the formulation of the model. Section (3) will analyze the results of simulations of monsoonal flow and section (4) will draw the conclusions.

2 The model

2.1 The averaged equations

Consider the following axially symmetric equations for a hydrostatic atmosphere:

$$\frac{\partial u}{\partial t}+\frac{v}{a^2\cos(\theta)}\frac{\partial M}{\partial \theta} = -\epsilon u \tag{1}$$

$$\frac{\partial v}{\partial t}+fu+\frac{u^2}{a}\tan(\theta) = -\frac{1}{a}\frac{\partial \Phi}{\partial \theta}-\epsilon v \tag{2}$$

$$\frac{\partial \Phi}{\partial \pi} = -C_p\Theta \tag{3}$$

$$\frac{\partial \dot{\pi}}{\partial \pi}+\frac{1}{a\cos(\theta)}\frac{\partial v\cos(\theta)}{\partial \theta} = 0 \tag{4}$$

$$\frac{\partial \Theta}{\partial t}+\frac{v}{a}\frac{\partial \Theta}{\partial \theta}+\dot{\pi}\frac{\partial \Theta}{\partial \pi} = Q \tag{5}$$

where $\pi=\left(\frac{p}{p_0}\right)^{R/C_p}$ is the Exner pressure, Θ is the potential temperature, Φ is the geopotential, $M=\Omega a^2\cos(\theta)^2+ua\cos(\theta)$ is the angular momentum density, $(u,v,\dot{\pi})$ is the 3D velocity vector, $f=2\Omega\sin(\theta)$ is the Coriolis parameter, θ is the latitude, ϵ is a Rayleigh friction, and Q is the diabatic heating. All variables are intended as large scale variables, resulting from an appropriate average which requires parameterization of organized convection.

Vertical advection of momentum and horizontal advection of meridional momentum have been ignored, based on the sole assumption that they are small for the flows that we are concerned with and therefore only quantitatively important for the problem.

To first approximation a monsoon such as the Asian monsoon is a meridional overturning circulation, with an upper branch uniformly flowing through most of the troposphere, the return flow being confined to a relatively thin surface mixed layer. We can therefore divide our model troposphere in two layers: one layer of thickness $\pi_{ML}-\pi_{top}$, representing the free atmosphere, overlaying a layer of thickness $\pi_0-\pi_{ML}$ representing the mixed layer. π_0, π_{ML}, π_{top} are the specified pressure heights of the surface, the interface between the mixed layer and the free atmosphere, and the top of the troposphere respectively.

Integrating equation (3) vertically we obtain for the two layers:

$$\begin{cases} \Phi_{ML}-\Phi_{top} = -C_p\int_{top}^{ML}\Theta d\pi \\ \Phi_0-\Phi_{ML} = -C_p\int_{ML}^{0}\Theta d\pi \end{cases} \tag{6}$$

so that the geopotential thickness is the vertically integrated temperature profile (for convenience we set $\Phi_0=0$).

If we impose

$$\pi_{ML}-\pi_{top}\gg\pi_0-\pi_{ML} \tag{7}$$

it is reasonable to assume that

$$\Phi(\pi_{top})\gg\Phi(\pi_{ML}). \tag{8}$$

We will also assume that
$$\frac{\partial \Phi\left(\pi_{top}\right)}{\partial \theta} \gg \frac{\partial \Phi\left(\pi_{ML}\right)}{\partial \theta} \tag{9}$$
which for an atmosphere close to geostrophy is equivalent to
$$u\left(\pi_{top}\right) \gg u\left(\pi_{ML}\right). \tag{10}$$

For zonal wind at the top of the troposphere the pressure force in equation (2) is then simply
$$\frac{1}{a}\frac{\partial \Phi_{top}}{\partial \theta} = \frac{g}{a}\frac{\partial h}{\partial \theta} \tag{11}$$
where
$$h = \frac{C_p}{g}\int_{\pi_{top}}^{\pi_{ML}} \Theta d\pi \tag{12}$$

In each layer the meridional velocity v is assumed constant, so that equations (1), (2) with the pressure force given by equation (11) are momentum equations for the meridional flow in the free atmosphere and zonal flow at the top of the troposphere. The system of equations must be closed with a theory for h.

2.2 The thermodynamic equation and the convective closure

Vertically integrating equation (5) in the free atmosphere gives:
$$\frac{\partial h}{\partial t} + \frac{v}{a}\frac{\partial h}{\partial \theta} + \frac{C_p}{g}\int \dot{\pi}\frac{\partial \Theta}{\partial \pi} d\pi = \frac{C_p}{g}\int Q d\pi \tag{13}$$

The third term on the l.h.s. of equation (13) can be integrated by parts to obtain
$$\int \dot{\pi}\frac{\partial \Theta}{\partial \pi} d\pi = \int \frac{\partial \dot{\pi}\Theta}{\partial \pi} d\pi - \int \frac{\partial \dot{\pi}}{\partial \pi}\Theta d\pi = \int \frac{\partial \dot{\pi}\Theta}{\partial \pi} d\pi + \frac{g}{C_p}\frac{h}{a\cos\left(\theta\right)}\frac{\partial v\cos\left(\theta\right)}{\partial \theta} \tag{14}$$
where the last equality has been derived using the continuity equation (4) and the fact that we have assumed meridional wind independent of height.

Assuming $\dot{\pi}_{top} = 0$ we obtain
$$\int \frac{\partial \dot{\pi}\Theta}{\partial \pi} d\pi = \dot{\pi}_{ML}\Theta_{ML} \tag{15}$$
where $\dot{\pi}_{ML}$ can be calculated from the continuity equation (4) to be
$$\dot{\pi}_{ML} = -\int_{\pi_{top}}^{\pi_{ML}} \frac{1}{a\cos\left(\theta\right)}\frac{\partial v\cos\left(\theta\right)}{\partial \theta} d\pi = -\frac{1}{a\cos\left(\theta\right)}\frac{\partial v\cos\left(\theta\right)}{\partial \theta}\left(\pi_{ML} - \pi_{top}\right) \tag{16}$$
Finally, substituting equations (14), (16) into (13) we get:
$$\frac{\partial h}{\partial t} + \frac{v}{a}\frac{\partial h}{\partial \theta} + \left(h - \frac{C_p}{g}\Theta_{ML}\left(\pi_{ML} - \pi_{top}\right)\right)\frac{1}{a\cos\left(\theta\right)}\frac{\partial v\cos\left(\theta\right)}{\partial \theta} = \frac{C_p}{g}\int Q d\pi \tag{17}$$
The term $\left(h - \frac{C_p}{g}\Theta_{ML}\left(\pi_{ML} - \pi_{top}\right)\right)$ is a measure of the dry static stability of the atmosphere, which varies as the integrated temperature changes: if the upper layer (*i.e.* the free

atmosphere) is heated up then h will increase and the amount of energy necessary to lift a unit thickness slab of atmosphere in a unit time will increase accordingly.

The diabatic forcing Q represents the effect of moist convection, radiative cooling and unresolved turbulent fluxes on the temperature of the free atmosphere. In the absence of large scale circulation ($v = 0$) radiative cooling and convection would compete, driving the atmosphere towards a state of radiative-convective equilibrium.

We will assume that such a state is defined by a specific thermal structure so that the value of CAPE is fixed to zero:

$$CAPE = \int_{p_{TOP}}^{p_0} (\alpha_p - \alpha_e)\, dp \simeq 0 \tag{18}$$

with α_p is the specific parcel volume and α_e is the specific volume of the environment, so that the atmosphere is neutrally buoyant [1].

We will then require that the model satisfy equation (18) wherever convection, whether dry or moist, occurs. This means that the temperature profile in the free atmosphere will be that deduced by the pseudo-adiabatic lifting of a parcel of air from the mixed layer: in the absence of large scale circulation, the clear atmosphere relaxes to radiative equilibrium on a timescale of about 10 days; this profile of temperature is unconditionally unstable for normal surface temperatures so that convection develops spontaneously; parcels therefore rise following a dry adiabat up to the lifting condensation level (LCL) whence they follow a pseudoadiabat up to where their buoyancy meets that of radiative equilibrium. Radiative cooling always drives back the atmosphere to an unstable profile, so, as the convective process is much faster than radiative cooling, the temperature profile in a convecting region should be close to that traced by a particle lifted pseudo-adiabatically.

We now assume that in the presence of convection all vertical mass flux originates in the mixed layer: knowledge of the temperature and moisture in the mixed layer is then sufficient to determine the vertical temperature structure of a convecting region.

It follows that we expect to substitute the r.h.s. of equation (17) with a relaxation towards the integrated temperature profile of a convecting atmosphere, calculated using the definition of equation (12). The problem has been reduced to determining the values of moisture and potential temperature in the mixed layer.

Axisymmetry requires that the meridional mass flow in the mixed layer be equal and opposite to that in the free atmosphere, so that at any latitude the meridional mass transport is zero. Therefore the velocity in the mixed layer is:

$$v_{ML} = \frac{\Delta\pi_{free}}{\Delta\pi_{ML}} v \tag{19}$$

A source of moisture is geographically specified by restoring (typically over the ocean) to saturation. The moisture budget is closed by including advection by the large scale flow as well as drying of the mixed layer by entrainment of dry air from above, so that the equation for moisture in the mixed layer is:

$$\frac{\partial q}{\partial t} + \frac{1}{a\cos(\theta)}\frac{\partial q v_{ML}\cos(\theta)}{\partial\theta} = \begin{cases} 0 & \text{if over land} \\ \frac{q_s - q}{\tau_m} & \text{if over ocean} \end{cases} \tag{20}$$

[1] There is some observational evidence that $CAPE \simeq 0$ might be the right assumption but the subject matter is controversial [4].

where q is the specific humidity of the mixed layer. The relaxation towards saturation q_s has a timescale τ_m of typically 1-2 days. All condensation is assumed to occur above the mixed layer and notice that the balance doesn't explicitly contain precipitation. q_s is a function of the temperature of the mixed layer alone, and therefore a function of the surface temperature, as the model is forced by a specified surface temperature, and we assume that the entire mixed layer has potential temperature equal to the imposed surface temperature.

Equation (20) is a crude approximation to the moisture budget. The model does not store soil moisture, so evaporation and precipitation over land are not part of the balance. This ensures that the ocean is the only moisture source for the model, a limitation that could be overcome in future work by allowing a more sophisticated model for the hydrology.

Convection will not always occur. Whenever the large scale circulation makes the free atmosphere more buoyant than convection would, convection must be inhibited. In this simplified model this occurs when h, the vertically integrated temperature of the free atmosphere, is greater than the vertically integrated temperature of a convecting atmosphere, with the same moist properties as the local mixed layer. In these non convecting regions the atmosphere must be relaxing towards radiative equilibrium with a timescale of about 10 days. This situation is equivalent to a region of trade inversion in the real atmosphere.

Given all the above, the following closure for moist convection is adopted:

$$\frac{C_p}{g}\int \overline{Q}d\pi = \frac{H_e - h}{\tau_h\left(\theta\right)} \tag{21}$$

with

$$\begin{cases} H_e &= \frac{C_p}{g}\int \Theta_p(\pi)d\pi, \quad \tau_h \simeq 6\,\text{hours} \quad \forall h < \frac{C_p}{g}\int \Theta_p(\pi)d\pi \\ \\ H_e &= \frac{C_p}{g}\int \Theta_r(\pi)d\pi, \quad \tau_h \simeq 10\,\text{days} \quad \forall h > \frac{C_p}{g}\int \Theta_p(\pi)d\pi \end{cases} \tag{22}$$

where $\Theta_p = \Theta_p(\pi)$ is the profile of potential temperature followed by a parcel rising pseudo-adiabatically from the mixed layer, and Θ_r is a radiative equilibrium potential temperature profile for the free atmosphere which we assume constant at $\Theta_r \simeq 220^oK$. Notice that the condition for convection amounts to a comparison of the dry static energy of the free atmosphere, as measured by the integrated temperature, with the moist static energy of the mixed layer. Therefore because moisture is not predicted in the free atmosphere the conditions more stringent than it should be.

No mention is explicitly given to large scale convergence or vertical velocity in the convective parameterization. Although precipitation is not thermodynamically coupled to the flow, potential for a feed-back between moist convergence and large scale circulation does exist, given the appropriate initial conditions, as will be seen in section (3.2)

The final equations for the free atmosphere then take the form:

$$\frac{\partial u}{\partial t} + \frac{v}{a^2 \cos(\theta)} \frac{\partial M}{\partial \theta} = -\epsilon u \tag{23}$$

$$\frac{\partial v}{\partial t} + fu + \frac{u^2}{a} \tan(\theta) = -\frac{g}{a} \frac{\partial h}{\partial \theta} - \epsilon v \tag{24}$$

$$\begin{aligned} \frac{\partial h}{\partial t} + \frac{v}{a} \frac{\partial h}{\partial \theta} + \left(h - \frac{C_p}{g} \Theta_{ML} \left(\pi_{ML} - \pi_{top} \right) \right) & \\ \cdot \left(\frac{1}{a \cos(\theta)} \frac{\partial v \cos(\theta)}{\partial \theta} \right) &= \frac{H_e - h}{\tau} \end{aligned} \tag{25}$$

τ is now a function of the convective state of the model, varying between a few hours and 10 days. H_e is the integral of the restoring temperature profile, which is either the specified radiative equilibrium, or the convective value, dependent on the local values of temperature and moisture in the mixed layer.

Two different processes besides advection can change the integrated temperature in equation (25): one is the term on the r.h.s. representing convection and physical processes, and the other is the last term on the l.h.s. representing the stabilization/destabilization of a convergent/divergent large scale circulation. We will return to the relative magnitude and importance of these two processes in the last section.

2.3 Steady state solutions in the limit of no viscosity

In the limit of small viscosity, the solution to equations (23), (24), (25) becomes independent of the actual numerical value of ϵ. A small but finite ϵ is nevertheless necessary for numerical stability, so a timescale of 400 days (much greater than any other timescale in the problem) is chosen for ϵ^{-1}.

Notice that Hide's theorem as stated in [3] cannot be exactly applied in this case. Equation (23) in steady state flux form is

$$\frac{\partial v M \cos(\theta)}{\partial \theta} = M \frac{\partial v \cos(\theta)}{\partial \theta} - \epsilon u a^2 \cos(\theta) \tag{26}$$

so that for a sufficiently small domain including a maximum of M the integral of the l.h.s. of equation (26) is zero. The integral of the r.h.s. is also zero if the local maximum of angular momentum density coincides with a divergent flow field (a mass source for the free atmosphere) so Hide's theorem does not hold and a maximum of M in the free atmosphere is possible. The problem arises from neglecting the vertical advection of momentum. However, a localized maximum of angular momentum density off the equator would be inertially unstable to small perturbations so that when integrating the time dependent equations such a solution would not occur. It has to be noted however that as formulated the model is not incompatible with equatorial superrotation.

Only steady state solutions of equations (23), (24), (25) will be considered here. As described in [1] two solutions exist for the zonal wind profile in the steady state inviscid limit: one in thermal wind equilibrium with the forcing and one angular momentum conserving. The former implies no meridional circulation while the latter implies a strong meridional circulation. Both are solutions of the approximate equation

$$\frac{\partial M}{\partial \theta} v \sim 0 \tag{27}$$

Accordingly, the steady state form of equation (25) can be satisfied in one of two ways:

$$i) \qquad h(\theta) = H_e(\theta) \tag{28}$$

$$ii) \qquad \int \left[\left(\frac{C_p}{g} \Theta_{ML} \left(\pi_{ML} - \pi_{top} \right) \right) \cdot \left(\frac{1}{a\cos(\theta)} \frac{\partial v \cos(\theta)}{\partial \theta} \right) + \frac{H_e - h}{\tau} \right] d\theta = 0 \tag{29}$$

equation (28) corresponding to $v = 0$, and equation (29) corresponding to an integral balance which determines the extent and strength of the meridional circulation.

Whether the solution is of the first or second type depends on the strength of the forcing. Suppose a distribution of H_e is imposed. The adjustment from rest will involve a progressive increase of h towards H_e. In doing so a pressure a pressure gradient will be established and a meridional circulation will start, which will then produce a zonal wind in geostrophic equilibrium with the pressure gradient. As h increases the flow goes through successive quasi-equilibrium states of geostrophy. If H_e does not involve too steep gradients, h eventually reaches H_e and the equilibrium solution is attained with no meridional flow. If on the other hand H_e does involve steep gradients, the meridional flow involved in the adjustment can be strong enough that the equilibrium expressed by equation (29) is reached before (28) is achieved and a steady state meridional circulation results, which will be angular momentum conserving so that $\frac{\partial M}{\partial \theta} = 0$.

A condition on H_e can be calculated to discriminate between the two solutions. The condition, an expression for which can be found in [1], states that a meridional circulation will be a steady state solution if the imposed H_e were to produce a geostrophycally adjusted zonal wind such that $\frac{\partial M}{\partial \theta} \geq 0$ in the northern hemisphere. This is equivalent to requiring that the flow be on the verge of inertial instability, although, as the adjustment sequence described above implies, inertial instability is actually never reached.

This description is inadequate for our model as H_e cannot be imposed: it depends instead on the moisture distribution in the mixed layer which is predicted by the model. We therefore turn to numerical integration.

3 The monsoon

The model described by equations (23), (24), (25), (20) with condition (22) was discretized on a staggered grid with meridional resolution of 200 points, corresponding to slightly less than one degree per grid box. The model was run to steady state using a leapfrog timestepping with Asselin filter.

3.1 Dry runs

The presence of moisture modifies the temperature structure of the convecting atmosphere, and the fact that moisture is interactive complicates the issue further. Before we analyze the behavior of the moist model, we can explore the dry model in which moist convection is turned off. In this case H_e is either the integral of the radiative equilibrium profile, or

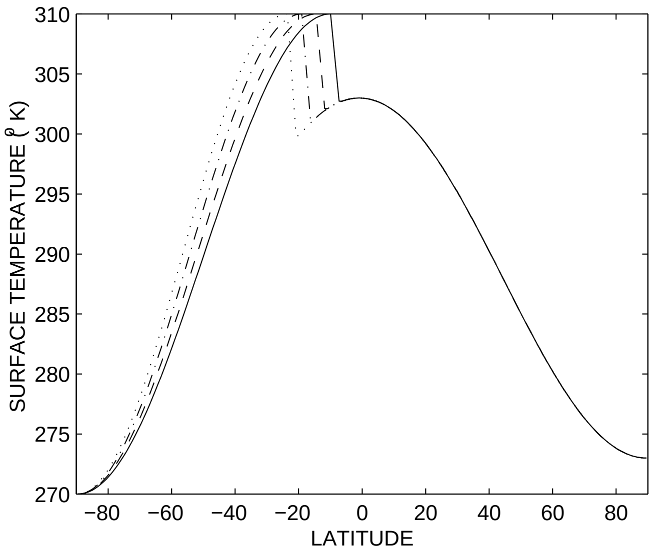

Figure 1: Surface temperature forcing. Two maxima are present: one representing the oceanic maximum and one the land. The maximum oceanic temperature is at the equator. The maximum land temperature is reached at 10^o, 15^o, 20^o, 25^o degrees latitude for the solid, dashed, dash-dotted, straight lines respectively.

the result of dry convection, which is only dependent on the potential temperature of the mixed layer and therefore on the imposed surface temperature.

Four profiles of surface temperature are shown in figure (1). The temperature profile exhibits two local maxima, one at the equator of about 30^oC, and one off the equator of 37^oC, which are chosen to represent the maximum oceanic temperature and the maximum land temperature respectively. There is no other distinction between land and ocean. Figure (2) shows the meridional profiles of zonal wind corresponding to the four cases of surface temperature. For all four the solution involves an angular momentum conserving meridional circulation, meaning that the meridional temperature gradient is sufficiently sharp to sustain a meridional circulation.

Two different behaviors are observed. For an off-equatorial heating sufficiently far from the equator, a local circulation is established. In these cases (bottom panel) the mass balance (29) is established within a few degrees of the forcing. Equatorward of that circulation an ordinary Hadley cell is established (the same case has been run without a temperature maximum at the equator and the same off equatorial circulation develops).

For simulations were the off-equatorial heating is closer to the equator, only one cross-equatorial circulation develops and the Hadley cell disappears (figure (2), top panel).

Two of these cases are compared in figure (3). Sinks of mass, corresponding to where the free atmosphere is not convecting and therefore cooled towards radiative equilibrium, are clearly represented by regions where the restoring height is much smaller than the simulated height. In case 1 the model develops a strong meridional circulation centered on the maximum temperature of the equator. In case 4 the off equatorial maximum is sufficiently far from the equator that an independent Hadley cell develops. The different

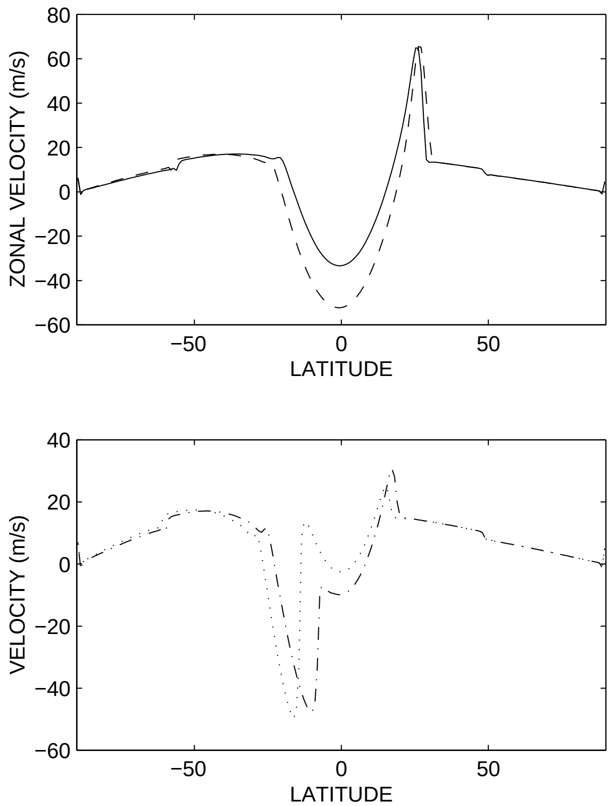

Figure 2: Zonal velocity. Top: cases 1 and 2 corresponding to a maximum continental temperature at 10^o(solid) and 15^o(dashed) latitude. Bottom: cases 3 and 4 corresponding to a maximum continental temperature at 20^o(dash-dotted) and 25^o(dotted) latitude.

circulation is particularly evident in the meridional wind profiles (figure (3), bottom panel). For case 1 we can see one large meridional circulation whereas for case 4 two cells are present.

Without moisture it appears to be very difficult not to obtain a meridional circulation. Temperature contrasts of a few degrees, distributed over a distance of a few degrees latitude, representing the temperature contrasts observed at the onset of the Asian monsoon, seem to be sufficient to obtain a meridional circulation. As the maximum surface temperature is moved away from the equator though, the cross-equatorial circulation is lost to a more local one; we will focus only on cross-equatorial circulations as models of the Asian monsoon.

The presence of moisture and maybe convection over the ocean must be crucial in preventing the formation of a monsoon, when India is hot but dry, as happens in May [2] just before the onset of the monsoon itself.

3.2 Moist runs

To understand the effects of moisture we first seek a distribution of surface temperatures that in the absence of moist convection would not be able to trigger a monsoon. In order to obtain such equilibrium solution, an unrealistic temperature distribution is needed (figure (4)): the maximum land temperature is placed at 30^o latitude off the equator and the transition to oceanic temperatures occurs over 25^o latitude. When moisture is turned off, the flow is in substantial equilibrium with the thermal forcing (figure (5), top panel), and only a hemispheric Hadley Cell is present.

Two moist cases are now presented. In the first case moisture is forced over the ocean only. No moisture is present over land but the one brought in by advection by the mean flow. This simulation represents the state of the Indian region before the onset of the monsoon, with a hot and dry Indian continent and a warm ocean. Land starts at 5^oS latitude and spreads southward to the pole.

As shown in figures (5), middle panel, and (6), the effect of moisture overrides the temperature gradient. As soon as the mixed layer is moistened, the convective temperature profile makes the equivalent height of the atmosphere over the ocean much greater than that over land. A strong meridional circulation then develops, with rising air over the ocean and subsiding elsewhere, in particular over land. In this situation a monsoon develops opposite to the observed.

The region of convection is very narrow and positioned off the equator on the hemisphere opposite to were land is. The width of the convecting region is the result of the cooperation of advection of moisture and the drying effect of upper tropospheric convergent flow, which tend to concentrate moisture in a delta-form distribution.

In the absence of a triggering mechanism, initial conditions similar to the observed are insufficient to initiate the development of a monsoon. If on the other hand all that is missing is in fact a triggering mechanism, if we artificially bring in some moisture over land, then the circulation should be as the one observed and should be able to sustain itself.

We therefore change the initial conditions by imposing that land be initially moist. If during the the initial part of the simulation we also allow land to be a source of moisture, then all the domain is moist and the temperature gradient establishes the direction of the circulation. If the moisture source over land is subsequently turned off, so that the forcing

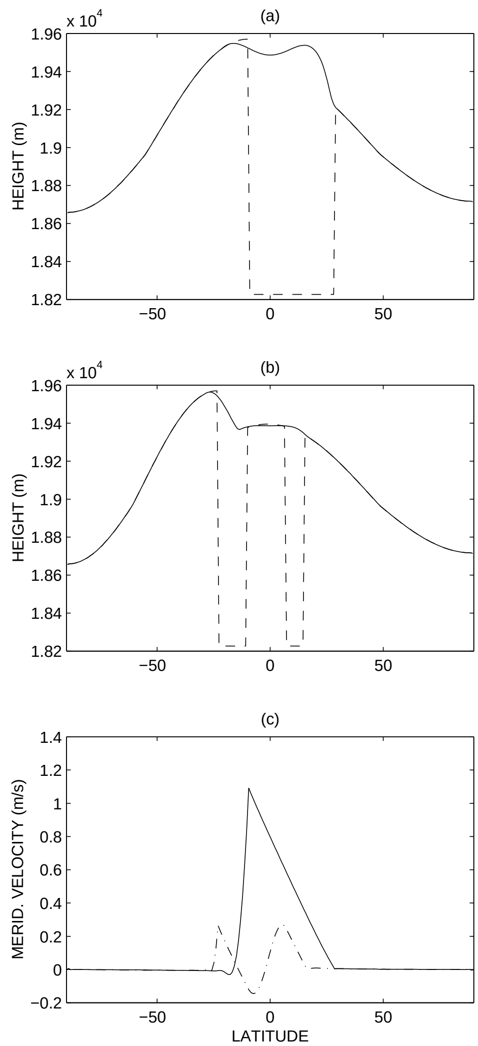

Figure 3: (a) Height (solid) and restoring height (dashed) for case 1 (10^o max. temperature). (b) Same as (a) but for case 4 (10^o max. temperature). (c) Meridional velocity for case 1 (solid) and 4 (dashed).

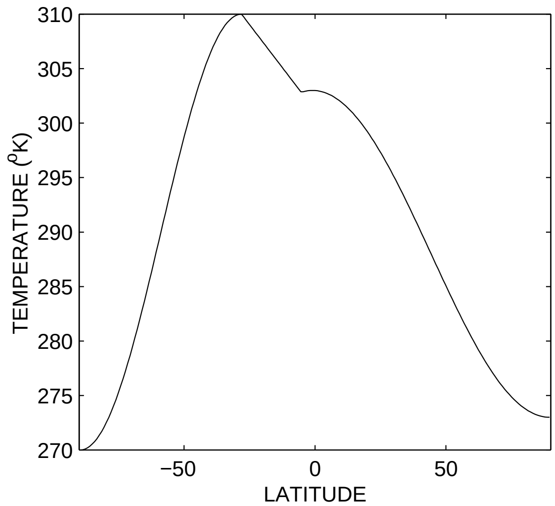

Figure 4: Temperature forcing for the moist experiment on Xie's arguments [2]

is now identical to the previous case, the circulation does not change and remains divergent over land (figure (5), bottom panel, figure (6)).

Once moisture is introduced over land, a circulation is established that is unable to dry out completely the mixed layer and draws in water from the ocean, which is now upwind, so that the circulation is able to sustain itself.

The ocean is, unrealistically, the only source of moisture in this model. Note that although heating is implied by the value of h, we have made no attempt to diagnose precipitation from that implied heating. Neglecting precipitation and subsequent evaporation from the ground is an important shortcoming of the model. Furthermore, drying of the mixed layer is accomplished through large scale convergence, whereas convective downdrafts, also diagnosed from the implied heating, would have been a better choice. These improvements are left for future work.

Despite these deficiencies, these simulations show that the same surface forcing with different initial conditions produces different solutions: the axisymmetric model exhibits multiple equilibria.

4 Discussion and conclusions

It has been argued [2] that a monsoon in a simplified geometry is the result of a zonally asymmetric disturbance that brings moisture over land. Before the disturbance propagates the wind profile is in thermal equilibrium with the surface forcing. The explosive nature of the monsoon is related to the abruptness of the instability that suddenly brings moisture over land, initiating moist convection.

An axially symmetric model was constructed to analyze these questions in a simplified setting. The formulation of this model differs from previous works in many ways: it is a model for the vertically integrated temperature profile and the upper tropospheric flow; it does not require the specification of a mean static stability; it contains a physically

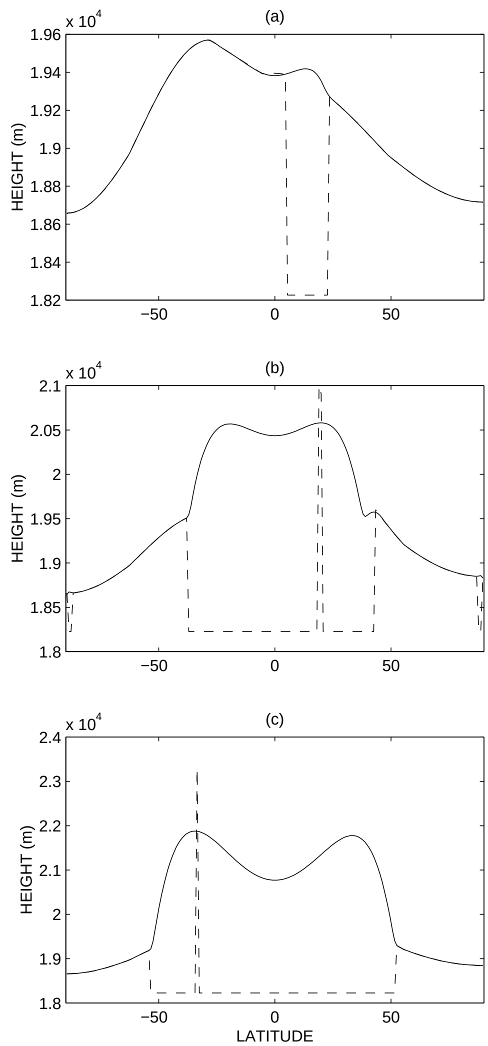

Figure 5: (a) Height and restoring height for the dry circulation. (b) Same as (a) but for moisture forced over ocean. (c) Same as (b) but with also initial moisture over the land added. Land starts where the temperature begins increasing poleward.

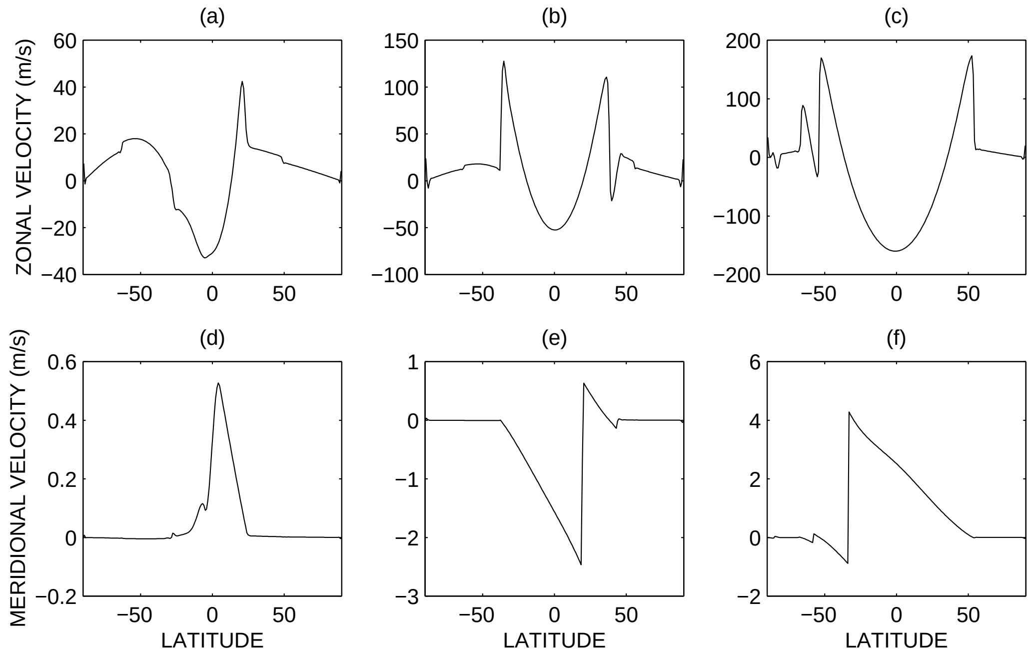

Figure 6: (a), (b), (c), zonal velocities for the dry case, the only ocean moist case and the moist land case respectively. (d), (e), (f) same as (a), (b), (c) but for meridional velocity.

motivated parameterization for moist convection.

In the absence of moist convection a thermal wind in balance with the forcing is hard to achieve unless the temperature contrasts are spread out over a few thousand kilometers. Starting from a state that is indeed in equilibrium without moist convection, we have shown that the final steady state solution is a function of the initial distribution of moisture.

Initializing the model with a dry and hot continent and a source of moisture over the ocean, as is the case for India before the onset of the monsoon, does not produce a monsoon, but rather a circulation with the same character but opposite in sign. If moisture is artificially introduced over land, then a monsoon like the observed develops. The model can therefore exhibit multiple equilibria.

The model does allow for the adjustment of static stability. We have seen in section (2.2) that there are two processes besides advection that can alter the temperature of the free atmosphere: convective processes and large scale convergence. In all experiments the convective closure is the dominant process heating (cooling in the case of non-convecting atmosphere) the free atmosphere. In all experiments, convergence in the presence of a positive static stability never grows to more than 5 % of the convective closure. Furthermore, variations of static stability in this model are limited to no more than 15 %, making convergence a secondary process from a thermodynamical point of view.

More work is needed to establish the behavior of this model in a wider parameter range: multiple equilibria have been achieved using a highly unrealistic surface temperature profile. For very shallow temperature gradients, moisture is necessary to trigger any kind of off equatorial meridional circulation. We have seen though that more realistic surface temperature profiles are able to trigger a monsoon in the complete absence of moisture. It is possible that by introducing moisture over the ocean such temperature profiles would still exhibit multiple equilibria, but for sufficiently strong temperature gradients we expect this behavior to disappear. What is the behavior of the model between these two extreme cases needs to be investigated.

Acknowledgements

I would like to thank Alan Plumb for suggesting this project and helping throughout; Isaac Held for an inspiring set of lectures and many useful discussions; Wayne Schubert for his infinite patience and for the many enlightening discussions on convection; Rick Salmon, for organizing an unforgettable summer; all the staff at Walsh Cottage, for being such a remarkable group. Finally, I would like to thank all my friends, Karen, Cheryl, Zhiming, Chris W, Christos, Lucy, Chris S, Tivon, and of course the Pontiac Catalina. Thank you all.

References

[1] R. A. Plumb and A. Y. Hou, "The response of a zonally symmetric atmosphere to subtropical thermal forcing," J. Atmos. Sci. **49**, 1790 (1992).

[2] S.-P. Xie and N. Saiki, "Abrupt onset and slow seasonal evolution of summer monsoon in an idealized gcm simulation," J. Met. Soc. Japan **77**, 949 (1999).

[3] I. M. Held and A. Y. Hou, "Nonlinear axially symmetric circulations in a nearly inviscid atmosphere," J. Atmos. Sci. **37**, 515 (1980).

[4] K. M. Xu and K. A. Emanuel, "Is the tropical atmosphere conditionally unstable?," Mon. Wea. Rev. **117**, 1471 (1989).