Monsoon dynamics with interactive forcing, Part I: Axisymmetric studies

Nikki C. Privé* and R. Alan Plumb

Program in Atmospheres, Oceans, and Climate,
Massachusetts Institute of Technology, Cambridge, Massachusetts

*Corresponding author address: N. Privé, 54-1717 MIT, 77 Massachusetts Ave., Cambridge, MA 02139.
email: nprive@alum.mit.edu
Abstract

The applicability of axisymmetric theory of angular momentum conserving circulations to the large-scale steady monsoon is studied in a general circulation model with idealized representations of continental geometry and simple physics. Results from an aquaplanet setup with localized subtropical forcing are compared with a continental case. It is found that the meridional circulation which develops is in good agreement with the nonlinear theory, both in the aquaplanet and in the continental cases. The equator proves to be a substantial barrier to boundary layer meridional flow; flow into the summer hemisphere from the winter hemisphere tends to occur in the free troposphere rather than in the boundary layer. A theory is proposed to explain the location of the monsoon; assuming quasi-equilibrium, the poleward boundary of the monsoon circulation is colocated with the maximum in subcloud moist static energy, with the monsoon rains occurring slightly equatorward of this maximum. The model results support this theory of monsoon location, and it is found that the subcloud moist static energy distribution is determined by a balance between surface forcing and advection by the large scale flow.
1. Introduction

This paper focuses on axisymmetric models of monsoon circulations as a first step towards developing an understanding of the large-scale circulation. The observed monsoon is strongly asymmetric, so that the applicability of a strict axisymmetric theory is questionable. The question of asymmetry of the flow will be addressed in a companion paper; the current work seeks to address only the purely axisymmetric case.

The classic view of the monsoon has been founded on the existence of a strong contrast in heating between the ocean and land, with the monsoon itself manifesting as an enormous sea-breeze (Halley, 1686). However, this depiction of monsoon dynamics fails to account for some of the observed behaviors of the monsoon, including abrupt delayed onset and the active-break cycle, and does not consider the impact of planetary rotation on such a large-scale flow. An alternative view which considers the monsoon as a seasonal displacement of the intertropical convergence zone (ITCZ) into the subtropics has recently gained support (eg. Chao and Chen (2001), Gadgil (2003)). It is this latter view which is the focus of this work.

The observed zonally averaged monsoon flow depicts a global meridional circulation cell with ascent in the monsoon region, outflow which crosses into the winter hemisphere aloft, subsidence in the winter hemisphere tropics, and cross-equatorial return flow at low levels. Using a linear shallow-water model, Gill (1980) found that a localized prescribed forcing in the off-equatorial tropics induces a cross-equatorial circulation similar to the observed monsoon flow. However, Held and Hou (1980), Lindzen and Hou (1988), and Plumb and Hou (1992) determined the axisymmetric Hadley circulation to be fundamentally nonlinear, and predicated upon the conservation of angular momentum in the free troposphere. One aim of this work is explore the validity of the nonlinear, axisymmetric theory of the steady Hadley circulation in describing the dynamics of the monsoon.

Held and Hou (1980) expanded upon the work of Hide (1969) and Schneider (1977) to explain the development of the annual mean Hadley cells using concepts of angular
momentum conservation. The predominance of a single Hadley cell in response to solstitial forcing was examined by Lindzen and Hou (1988). Plumb and Hou (1992) explored the atmospheric response to a localized subtropical forcing in a dry, axisymmetric framework. The critical condition for the development of an angular momentum conserving (AMC) meridional circulation is the existence of an extremum of angular momentum in the thermal equilibrium state. Emanuel (1995) showed that assuming a moist adiabatic lapse rate, and making use of the Maxwell relations, the threshold may be written as

\[
\left[ \frac{\partial}{\partial \phi} \left( \frac{\cos^3 \phi}{\sin \phi} (T_s - T_t) \frac{\partial s_b}{\partial \phi} \right) \right] = -4\Omega^2 a^2 \cos^3 \phi \sin \phi
\]  

(1)

where \( \phi \) is the latitude, \( a \) is the radius of the earth, \( \Omega \) is the angular velocity of the earth, \( s_b \) is the subcloud moist entropy, \( T_s \) is the surface temperature, and \( T_t \) is the temperature at the tropopause. Zheng (1998) verified threshold behavior in a moist axisymmetric aquaplanet model with fixed, local subtropical SST perturbation.

There has been interest in the application of the theory of threshold behavior to explain certain aspects of the monsoon. Plumb and Hou hypothesized that the abrupt onset of the monsoon might be related to this threshold behavior. Eltahir and Gong (1996) found that the subtropical meridional gradient of subcloud moist entropy was positively correlated with the strength of the West African monsoon.

There are several limitations to the existing nonlinear theory of the Hadley circulation. First, the work of Held and Hou (1980), Lindzen and Hou (1988), and Plumb and Hou (1992) is in a dry framework with an assigned distribution of radiative equilibrium temperature. In these cases, the induced circulation does not affect the forcing field, while in the real world, the forcing is highly dependent upon the circulation. For the moist experiments of Emanuel (1995) and Zheng (1998), a prescribed SST perturbation was used to force the atmosphere, with radiative convective equilibrium temperatures following a moist adiabat to communicate the surface forcing throughout the troposphere. Over a dry landmass forced by surface fluxes, the vertical column follows close to a dry adiabat, and upper level temperatures may
be relatively cold even though the lower tropospheric temperatures are high. This raises the question of whether strong sensible heat fluxes over an arid continent are sufficient to induce a monsoonal circulation. Also, given the interaction possible between the circulation and the forcing, the location and extent of the monsoon are not predictable from the radiative convective equilibrium state.

A second limitation is that the nonlinear theory is concerned with the steady state circulation, rather than the transient monsoon. The timescale for circulations to reach a steady state in axisymmetric models is frequently quite long, on the order of hundreds of days, which is much greater than the seasonal timescale associated with monsoons. Fang and Tung (1999) found that the abrupt increase in circulation strength observed when the steady forcing is shifted off of the equator was not seen when transient forcing was used.

In order to address the applicability of the nonlinear axisymmetric theory of Hadley circulations to the interactive monsoon, we wish to address the following questions:

1. How does the presence of a subtropical continent with interactive forcing affect the monsoon circulation?

2. What determines the location and extent of the monsoon?

3. Is the steady monsoon circulation representative of the dynamics of the transient monsoon?

An axisymmetric general circulation model is used to explore these questions.

There is a wide gap in modeling the monsoon between the highly idealized axisymmetric theory and full GCM studies with realistic physics. While the axisymmetric theory is useful for developing an understanding of the basic physical mechanisms which drive and affect the monsoon, it is unclear how the simplifications which are involved limit the applicability to the monsoon. On the other hand, the wealth of feedbacks present in the full GCM studies make diagnosis of the monsoon behavior extremely difficult. The goal of this work is to bridge the gap between the idealized axisymmetric theory and the more complex,
interactive monsoon. A general circulation model with simplified representations of some physical processes and with idealized continental geometry is chosen to achieve intermediate complexity, as described in Section 2. This allows for a reasonably more realistic portrayal of processes which are suspected to be intrinsic to the monsoon, while at the same time reducing the feedbacks to make analysis more tractable.

The first step is to characterize the Hadley response to a steady local subtropical forcing in an aquaplanet setup. The aquaplanet cases act as a basis of comparison for the later, more complex, cases with a subtropical continent. The subtropical forcing takes the form of an SST perturbation, the form of which is designed to mimic the presence of a landmass in order to allow direct comparison with continental experiments. The results of the aquaplanet cases are described in Section 3, where it is found that threshold behavior of the meridional circulation is seen as predicted by Plumb and Hou (1992) and Emanuel (1995). It is found that the strength of the circulation weakens as the forcing is shifted poleward.

Next, a simple subtropical continent is introduced with perpetual summer forcing, and comparison with the aquaplanet experiments helps to determine the impact of continental physics on the monsoon circulation. These experiments are discussed in Section 4. Threshold behavior of the circulation strength is not observed as clearly as in the aquaplanet cases, although the circulation does show a transition from local to global extent, as predicted by the nonlinear theory. A theory of monsoon location is introduced in Section 5. The boundary layer thermodynamics are shown to control the extent and location of the monsoon region such that the zero line of the circulation must be coincident with the maximum in subcloud moist static energy. Over the ocean, the moist static energy is closely related to the surface temperature, while over a land surface, the moist static energy is controlled by a balance between advection by the large-scale flow and surface heat fluxes.

Finally, seasonally varying forcing is implemented over the landmass to contrast with the perpetual summer cases to explore the applicability of the steady solutions to the transient monsoon. Section 6 addresses these experiments. The transient response approaches the
perpetual summer circulation by mid to late summer, but the early summer state is not close to the steady result. The timescale for the transient response is that needed for the large-scale overturning circulation to fold over the contours of angular momentum across the tropical upper troposphere. The overall findings are discussed in Section 7.

2. Model

The model used is the MIT General Circulation Model (MITGCM), release 1.0. The MITGCM consists of a dynamical core coupled to an atmospheric physics package; the dynamical kernel of the model is described by Marshall et al. (2004). The atmospheric MITGCM has been tested extensively against Held and Suarez (1994), although a different atmospheric physics package is implemented here. The model gridspace used is a partial sphere between 64S and 64N, with 40 pressure levels in the vertical at 25 mb intervals. A staggered spherical polar grid is used with 4° latitudinal resolution. There is no orography, and the surface is assigned to the 1012.5 mb pressure level. The coefficient of vertical viscosity is $10 \, Pa^2/s$, and an eighth-order Shapiro filter is employed to reduce horizontal noise in the temperature, humidity, and horizontal flow fields.

Radiation and cloud physics are not included; instead, the atmosphere undergoes Newtonian cooling with a timescale of $\tau_{NC} = 60$ days:

$$Q_{NC} = \tau_{NC}^{-1}(T_{NC} - T)$$

(2)

where $T$ is the temperature at a gridpoint, and $Q_{NC}$ is the cooling rate. $T_{NC} = 200$ K for all gridpoints; this profile for $T_{NC}$ is chosen for its simplicity.

The moist convective scheme of Emanuel (1991) is used, including the modifications of Emanuel and Živković-Rothman (1999). The convective parameters used as part of this scheme have been optimized against observed data from the Tropical Ocean Global Atmosphere Coupled Ocean-Atmosphere Response Experiment (Emanuel and Živković-Rothman,
The convection scheme includes dry adiabatic adjustment, which is performed over regions which are unstable to unsaturated ascent. A mixed layer of momentum is included at the lowest 200 mb of the model; in this layer, horizontal velocities are homogenized over a timescale of 500 sec.

Very simple representations of ocean and land surfaces are implemented. SSTs are prescribed over the ocean, with temperature profile fixed in time. Over land, a bucket hydrology following Manabe (1969) is used. Surface evaporation is modified by a factor $B$

$$B = \begin{cases} 
1 & B \geq 0.75B_0 \\
\frac{B}{0.75B_0} & B < 0.75B_0 
\end{cases} \quad (3)$$

where $B_0$ is an assigned bucket depth (20 cm) indicating the amount of moisture that can be stored per unit surface area, $B$ is the current moisture in the bucket per unit area, $E$ is the evaporation rate, and $P$ is the precipitation rate. Any excess moisture gained by precipitation is considered to be runoff. The initial moisture content of the buckets for each model case was assigned to be zero.

Because a Newtonian cooling scheme is used, the flux balance at the surface cannot be calculated using radiative fluxes. Instead, the net downward flux into the surface (THF) is prescribed as a function of latitude:

$$THF(\phi) = LHF(T_s) + SHF(T_s) \quad (4)$$

where $\phi$ is the latitude, LHF is the latent heat flux into the atmosphere, SHF is the sensible heat flux into the atmosphere, and $T_s$ is the surface temperature. The surface temperature and heat fluxes are interdependent and calculated iteratively. Using a prescribed net radiative flux has the benefits of allowing direct control over the land surface forcing and of reducing the number of feedbacks, such as cloud-radiative feedbacks, in comparison to a sit-
ulation with interactive radiation. This permits easier diagnosis of the underlying dynamical mechanisms, but at the cost of making the resulting flow less realistic. For example, surface temperatures may become very hot over a desert-like land area as there no increase in the outgoing longwave radiation with increased surface temperatures. Direct control of the land surface forcing allows the behavior of the monsoon to be explored in parameter space by testing a range of different forcing strengths.

3. Aquaplanet

The response of the atmosphere to a localized subtropical forcing is examined in an aquaplanet setup. The model is spun up from rest with SST of 302K at all latitudes for 200 days, then an SST perturbation of the form

\[ SST(\phi) = 302K + \Delta T \cos^2((\phi - \phi_0)\frac{5}{2}), \phi_0 < \phi < (\phi_0 + 36^\circ) \]

\[ SST(\phi) = 302K \ \phi \leq \phi_0, (\phi_0 + 36^\circ) \leq \phi \]

is introduced, where \( \Delta T \) is the strength of the SST perturbation, and \( \phi_0 \) is the location of the SST perturbation (Figure 1). The model is then integrated until an equilibrium state is reached. This form of the local SST perturbation is chosen to emulate the presence of a continent, with an abrupt interface between land and ocean in the subtropics. The uniformly warm ocean temperature profile is chosen to isolate the atmospheric response to the local SST perturbation. A range of \( \Delta T \) are tested to characterize any threshold behavior.

1. Subtropical Threshold Behavior

Two criteria are used to determine whether the modeled Hadley circulations are in agreement with the nonlinear theory: 1) conservation of angular momentum across the upper branch of the circulation cell; 2) existence of threshold behavior of the circulation strength as described
by Plumb and Hou (1992) and Emanuel (1995). The ocean forcing is located at $\phi_0 = 16N$ (5); this latitude is chosen as representative of a subtropical monsoon. The strength of the applied SST perturbation ($\Delta T$ in (5)) is varied from 0.5 $K$ to 2.5 $K$.

Threshold behavior of the circulation strength is clearly observed (Figure 2). When the SST perturbation is small, the resulting circulation is weak, and the upper tropospheric absolute vorticity does not approach the critical value at zero. Above the threshold forcing at $\Delta T = 1.25K$, the circulation intensifies much more rapidly with increased $\Delta T$, and the upper level absolute vorticity is close to zero. When forcing levels are below the threshold, the circulation is confined to one hemisphere and does not cross the equator (Figure 3a). Above the threshold forcing, the circulation becomes cross-equatorial and considerably stronger (Figure 3b). In these cases, the upper tropospheric absolute vorticity is close to zero in the circulation cell (Figure 2b), which is thus closely angular momentum conserving.

In the aquaplanet setup, there are strong feedbacks between the circulation and the surface fluxes, especially the latent heat flux. As the circulation intensifies, the surface winds increase, which enhances the surface heat fluxes and in turn strengthens the circulation; this type of interaction has been coined the ‘wind-induced surface heat exchange’ (WISHE) feedback (Emanuel, 1986). Threshold behavior might be exaggerated by WISHE, which would tend to strengthen already strong circulations but has less impact on weak circulations.

2. Cross-Equatorial Flow

There is a tendency for cross-equatorial circulations to ‘jump’ in the lower troposphere when approaching the equator (eg. Figure 3b). While some of the flow crosses the equator in the free troposphere, a portion of the circulation is confined to the southern hemisphere, giving the streamfunction the appearance of two conjoined Hadley cells. This flow pattern results in a secondary precipitation maximum in the southern hemisphere tropics between the equator and 6S. The moisture content of the low-level flow is depleted during the jump as the air rises to the mid-troposphere, but is replenished through large latent heat fluxes at the surface in
the northern hemisphere. Jumping also alters the streamfunction intensity. When jumping does not occur, the maximum streamfunction of the cross-equatorial circulation is located in the lower troposphere near the equator. The initiation of jumping eliminates the lower tropospheric streamfunction maximum, and the circulation maximum occurs in the upper troposphere.

Jumping of the meridional circulation has been addressed extensively by Pauluis (2001); a brief synopsis is given here. A pressure gradient across the equator is needed in the mixed layer in order to allow cross-equatorial flow. When the mixed layer is thin or the pressure gradient is weak, only a limited mass flux is possible in the mixed layer, so flow attempting to cross the equator must rise into the free troposphere in order to cross. In a moist atmosphere, the vertical moist stability near the equator is weak, and the resulting jump is quite deep. In the modeled case with uniform ocean SSTs, the pressure gradient across the equator is very weak, sometimes even increasing northward, so that cross-equatorial flow is strongly inhibited even though the mixed layer is 200 mb deep.

3. Latitudinal Influence on Circulation

A series of aquaplanet runs are made with uniform SST and a localized boreal hemisphere SST perturbation, which is varied in location. The form of the perturbation is given by 5, with $\phi_0$ varied from 0N to 26N, and $\Delta T$ is chosen to be 4.25 K. This value of $\Delta T$ was chosen to yield cross-equatorial circulations for the entire range of $\phi_0$ tested.

The impact of the location of the forcing on the strength of the circulation is illustrated in Figure 4. The maximum streamfunction increases as $\phi_0$ is moved poleward from 0N to 6N, although the circulation width only broadens slightly, with no sign of jumping behavior. As $\phi_0$ moves from 6N to 10N, the circulation nearly doubles in width but decreases sharply in strength as jumping initiates and the lower tropospheric streamfunction maximum is destroyed. As $\phi_0$ is moved poleward from 10N to 22N in individual steady state cases, the circulation decreases in strength and broadens slightly. As the circulation broadens,
the easterly jet intensifies significantly, resulting in greater viscous effects with weakening angular momentum conservation in the upper branch of the circulation. This increased viscosity, peculiar to numerical models, is hypothesized to account for the gradual weakening of the circulation as the forcing is shifted poleward. The $\phi_0 = 24N$ at $\Delta T = 4.25K$ case is close to the local-to-global AMC transition as described by Schneider (1983), but far above the critical threshold for the existence of an AMC circulation (Plumb and Hou, 1992). Schneider showed that an angular momentum conserving circulation forced by a $\delta$-function source undergoes a transition from a cell of regional extent to a cross-equatorial cell as the forcing is increased; the transition forcing needed is greater when the forcing is moved poleward. The proximity to this transition is the likely cause of the sharp reduction in streamfunction magnitude from $\phi_0 = 22N$ to $24N$.

4. Continental Cases

A subtropical continent with interactive surface temperatures and heat fluxes is introduced. The strength of the land surface forcing may be directly manipulated through the total surface heat flux (4), given by

$$ THF(\phi) = THF_0 - \Delta THF \sin((\phi - \phi_T))^2 $$

(6)

where $\Delta THF$ is 50 $W/m^2$, and $THF_0$ ranges from 120 to 150 $W/m^2$. For all of these cases, the continent extends from the southern coastline at $\phi_L$ northward to the model boundary at 64N. For each case, the model is first spun up from rest with a relatively cold continent ($THF_0 = 80W/m^2$) for 200 days. The land surface forcing is then increased to the summer value over a timescale of 100 days, and the model is run for at least 700 additional days until a steady state is reached.

First, a subtropical continent is used with $\phi_L = 16N$ and with $\phi_T$ of 8N in (6); the ocean is assigned uniform temperature. The only difference between this case and the previous
aquaplanet case with localized SST perturbation at 16N is the replacement of the prescribed SST perturbation with an interactive continent in the boreal hemisphere.

For the weakest continental forcing tested, $THF_0 = 120 W/m^2$, no monsoon occurs and there is little precipitation over the continent (not shown). The meridional circulation is limited to a very weak, shallow circulation just along the coastline, with subsidence in the mid and upper troposphere over the continental subtropics.

When the land forcing is increased to $125 \leq THF_0 \leq 130 W/m^2$, deep moist convection onsets over the subtropical continent. Large scale ascent occurs along the coastline, with a latitudinally narrow meridional circulation with subsidence over the tropical ocean (Figure 5). The upper tropospheric absolute vorticity (dotted line with squares, Figure 6b) is not close to the critical value at zero, and the deep circulation is not strongly angular momentum conserving.

For land forcing of $THF_0 \geq 135 W/m^2$, the meridional circulation is more global in extent, with ascent over the subtropical continent and subsidence over the southern hemisphere ocean (Figure 5b). Jumping of the circulation at the equator occurs for all cases with cross-equatorial flow, and becomes more pronounced as the land surface forcing increases. The upper level tropical easterly jet (not shown) is very strong, which is a common feature in axisymmetric models. The angular momentum field (not shown) is significantly distorted by the circulation, although the flow is able to cross some contours of angular momentum in the tropics, where there is strong easterly shear. As the land forcing is increased, the circulation broadens and the monsoon region moves inland, as shown in Figure 7.

Although the strength of the meridional circulation increases systematically with increased land surface forcing (dotted line with squares in Figure 6a), there is little indication of threshold behavior. The upper level absolute vorticity (Figure 6b) gradually approaches the critical value at zero for $THF_0 \geq 140 W/m^2$, but a cross-equatorial circulation develops when the vorticity is still sub-critical. The threshold behavior may be compared with that of the aquaplanet case in Figure 2.
Why does the aquaplanet case show clear threshold behavior while the continental case does not? There are two factors which contribute to this difference in behavior. First, the threshold behavior in the aquaplanet case is accentuated by WISHE-type feedbacks, as already discussed; however, these feedbacks do not occur over the continent given the constrained surface forcing. Second, the location of large scale ascent moves poleward with increased surface forcing (Figure 7) in the continental case, but is nearly stationary in the aquaplanet case. As shown previously (Figure 4), the circulation tends to weaken as the ascent region moves poleward through the subtropics; this would act to obscure threshold behavior in the continental case.

As in the aquaplanet cases, there is a transitional period while a steady circulation is established which ranges in duration from 200 to 250 days; this period shortens with increasing land forcing. After 50-100 days, the lower tropospheric leg of the meridional cell begins to show jumping behavior, which then becomes more severe as time progresses.

1. Continental Location

The location of the coastline is varied to further explore the applicability of the nonlinear theory to the monsoon. According to the theory and the results of the aquaplanet cases, a monsoon should require less forcing when the coastline is moved equatorward, and increased forcing as the coastline is moved poleward. A summer-like SST profile is implemented with subtropical continent in an effort to achieve more realistic flow in the equatorial region. The SST distribution is fixed and does not vary with time:

\[ SST(\phi) = SST_0 - \Delta T \sin(\phi - 8N)^2 \]  

(7)

where \( SST_0 = 302\,K \) and \( \Delta T = 28\,K \). Three coastline positions are tested: \( \phi_L = 16N, 8N, \) and \( 24N \). For each case, the surface forcing profile is shifted to match the coastline, with \( \phi_T \) in (6) 8° equatorward of \( \phi_L \), so that the magnitude of the forcing can be compared directly
as \( \phi_L \) is varied.

First, an aquaplanet case with the summer-like SST profile (7) is performed. In this case, the cross-equatorial ‘winter’ Hadley cell dominates with a much weaker ‘summer’ cell confined to the warmer hemisphere poleward of the SST maximum (Figure 8a). The Hadley cells nearly conserve angular momentum. The ascent region and the precipitation maximum (not shown) are located slightly equatorward of the SST maximum.

When a subtropical continent is added with this summer-like SST distribution, the overall meridional circulation is visually similar in appearance to a simple superposition of the flow from the summer-like aquaplanet case (Figure 8a) and the previous continental cases with uniform SST (Figure 5). Over the tropical ocean, a strong meridional cell forms as a result of the SST distribution: ascent occurs in the boreal hemisphere near the SST maximum (Figure 8b).

The case with \( \phi_L = 16\text{N} \) may be compared with the previous continental cases with uniformly warm ocean. For weak land forcing \((THF_0 \leq 135 \text{ W/m}^2)\), the streamfunction over the continent is almost twice as strong in the summer SST case as in the uniform SST case (dot-dash line with asterisks in Figure 6a), but for \( THF_0 \geq 140 \text{ W/m}^2 \), the circulation is similar in intensity. The 150 mb absolute vorticity over the continent is near the critical value of zero for \( THF_0 \geq 135 \text{ W/m}^2 \) (dot-dash line with asterisks in Figure 6b); however, there is no sign of threshold behavior of the circulation strength.

The coastline is relocated equatorward, at \( \phi_L = 8\text{N} \), with \( \phi_T = 0\text{N} \) in (6). With weak applied land surface forcing \((THF_0 \leq 125 \text{ W/m}^2)\), the circulation of the \( \phi_L = 8\text{N} \) case is very similar to that of the \( \phi_L = 16\text{N} \) coastline case (solid line with circles, Figure 6a). As the land forcing is increased further \((THF_0 \geq 130 \text{ W/m}^2)\), the continental circulation strength and precipitation become more intense in the \( \phi_L = 8\text{N} \) case compared to the \( \phi_L = 16\text{N} \) case. The flow does not show the strong jumping tendency seen in the \( \phi_L = 16\text{N} \) case, and there is significant mass transport in the boundary layer at the equator (Figure 8c). Because the coastline and the heated continent are close to the equator, the large-scale pressure gradient
across the equator in the mixed layer has a stronger negative-northward magnitude. The continental precipitation maximum is shifted 4° equatorward relative to the \( \phi_L = 16N \) case (not shown).

Several factors may contribute to the strengthening of the continental streamfunction for \( \phi_L = 8N \) in comparison to the \( \phi_L = 16N \) case. First, the region of large scale ascent is located closer to the equator, which has been shown in the aquaplanet case to yield more intense circulations. Second, jumping behavior significantly weakens the circulation in the \( \phi_L = 16N \) case. As the jumping tendency is not as strong in the \( \phi_L = 8N \) case, the circulation is not as severely affected. There is again little sign of threshold behavior of the circulation strength or the precipitation field.

Finally, the coastline is moved poleward to \( \phi_L = 24N \), with forcing shifted to \( \phi_T = 16N \) in Equ. (6). As the surface forcing is increased, the meridional circulation over the land becomes deep, but remains narrow in latitudinal extent and weak in comparison to the cross-equatorial Hadley cell (Figure 8d). The continental circulation cell is completely separate from the cross-equatorial cell for all land surface forcing levels. Although the strength of the continental cell (dashed lines with triangles in Figure 6a) is much less than that seen in the \( \phi_L = 16N \) cases, the net precipitation over the landmass is comparable. Examination of the absolute angular momentum field for the strong land forcing cases shows that although the continental circulation cell is weak and narrow, it is nearly angular momentum conserving in the upper troposphere. This type of local AMC cell is described by Schneider (1977), and is different from the local viscous regime described by Plumb and Hou (1992). In order for an AMC cross-equatorial circulation to form with ascent over the continent in the 24N case, the upper level easterly jet would be unreasonably strong at the equator, with wind speeds on the order of 150 m/s.
5. Theory of monsoon location

What determines the location of the monsoon precipitation and the size of the meridional circulation cell? Why does the monsoon tend to shift poleward with increased land forcing? With the help of a few assumptions, an extension of existing axisymmetric theory can explain much of the large-scale dynamics.

Emanuel et al. (1994) have shown that with a statistical equilibrium approach, the zonal wind field is closely related to the distribution of subcloud moist entropy. The assumption is made that in the vicinity of the deep convection, where there is large scale ascent, the vertical thermodynamic profile approaches a moist adiabat. As a result, the upper tropospheric virtual temperature field is strongly tied to the subcloud moist static energy near the monsoon. The free tropospheric zonal wind field is in thermal wind balance with the density field,

\[
\frac{\partial u}{\partial p} = \frac{1}{f} \frac{\partial \alpha}{\partial y}
\]

(8)

where \(\alpha\) is the specific volume and \(f\) is the Coriolis parameter. Maxwell’s equations relate the density to the moist entropy:

\[
\left( \frac{\partial \alpha}{\partial y} \right)_p = \left( \frac{\partial T}{\partial p} \right)_{s^*} \frac{\partial s^*}{\partial y}
\]

(9)

where \(s^*\) is the saturation moist entropy. By substituting (9) into (8), and since \(s^*\) is nearly constant in height with the assumption of a moist adiabatic lapse rate,

\[
s^* = s_b
\]


\[
\frac{\partial u}{\partial p} = \frac{1}{f} \left( \frac{\partial T}{\partial p} \right)_{s^*} \frac{\partial s_b}{\partial y}
\]

(11)
where \( s_b \) is the subcloud moist entropy. The subcloud moist entropy is closely related to the subcloud moist static energy, \( h_b \), where

\[
h = L_v q + (c_{pd} + r_t c_t) T + (1 + r_t) g z
\]

(12)

\[
\delta h_b \approx T_b \delta s_b
\]

(13)

with \( L_v \) the latent heat of vaporization, \( T_b \) the mean subcloud temperature, specific humidity \( q \), specific heat of dry air \( c_{pd} \) and of liquid water \( c_t \), mixing ratio of water vapor \( r \) and of total water content \( r_t \), and geopotential height \( z \). Thus, the zonal wind shear can be rewritten

\[
\frac{\partial u}{\partial p} \approx \frac{1}{f} \left( \frac{\partial T}{\partial p} \right)_s \frac{1}{T_b} \frac{\partial h_b}{\partial y}
\]

(14)

The summer hemisphere poleward limit, or boundary, of the cross-equatorial meridional circulation is seen in the previous model cases to be a zero streamfunction contour which is nearly vertical. Assuming a meridional circulation which conserves absolute angular momentum in the free troposphere, the circulation boundary must be located in a region of zero vertical wind shear. This can be seen by considering the vertical distribution of momentum as the boundary is approached from the monsoon region: at the boundary, there is flow into the column only in the boundary layer and out of the column only in the upper troposphere, with no other sources of momentum advection. Since there is net ascent in the monsoon region, the momentum must be constant throughout the vertical column in the troposphere.

The statistical equilibrium theory may be applied to this framework in order to tie the monsoon location to the subcloud moist static energy. Let us assume a forcing region in the subtropics which results in a localized area of high subcloud moist static energy. For the sake of this argument, the forcing region is considered to be sufficiently strong to meet the threshold criteria for creation of an angular momentum conserving meridional circulation (Plumb and Hou (1992), Emanuel (1995)). From (14), \( \frac{\partial u}{\partial p} = 0 \) when either \( \frac{\partial T}{\partial p} \) or \( \frac{\partial h_b}{\partial y} \) are zero. The vertical temperature gradient is nonzero, so the zero wind shear line, and thus the
circulation boundary, will occur at the latitude at which \( h_b \) is maximum (\( \frac{\partial h_b}{\partial y} = 0 \)). Large scale ascent will occur near and equatorward of this maximum in \( h_b \), with the boundary of the circulation coincident with the \( h_b \) maximum. The precipitation is colocated with the large scale ascent, so that the subcloud moist static energy distribution determines the location of the monsoon circulation and precipitation.

This theory can be tested with the model. An example of a monsoonal case showing the relationship between the circulation, zonal wind field, precipitation, and subcloud moist static energy is shown in Figure 12. In the monsoon region, the zonal wind field is weak, with near zero shear as required by the theory. The location of the poleward boundary of the monsoonal circulation cell is found to be coincident with the latitude of maximum subcloud moist static energy, with the precipitation peak occurring at or equatorward of this point. This correspondence between moist static energy and the monsoon location is found to hold in all of the modeled cases with a continent. However, in some of the aquaplanet cases, the ascent branch of the circulation is not closely AMC due to numerical filtering of the zonal wind by the Shapiro filter; in these cases the maximum subcloud \( h_b \) is located slightly equatorward of the boundary of the meridional circulation.

What, then, determines the subcloud moist static energy distribution? In radiative convective equilibrium with this model setup, the subcloud moist static energy profile follows that of the prescribed continental net surface heat flux \( THF(\phi) \), although the actual value of \( h_b \) is constrained by moisture availability. In radiative convective equilibrium, the greatest subcloud moist static energy occurs where the net surface fluxes are largest, over the coastal continent. Once the meridional circulation develops, the low-level flow carries air from the oceans, which has lower subcloud moist static energy, over the coastal regions, locally reducing the moist static energy so that the maximum energy is located inland. As the forcing is increased, the circulation also intensifies, with a greater flux of low moist static energy air being carried onto the continent which must be heated by the surface forcing to bring it to the RCE energy state (Figure 9). The inflowing air must travel further over
the landmass while being heated from the surface to reach the maximum subcloud moist static energy. The steady solution is formed by a balance between the various tendencies of subcloud moist static energy.

The theory also helps to explain the flow seen in the case with coastline at \( \phi_L = 8^\circ \); although the continent is shifted \( 8^\circ \) equatorward from the \( \phi_L = 16^\circ \) case, the monsoon location only shifts \( 4^\circ \) equatorward. The subcloud moist static energy profiles reveal that in the \( 16^\circ \) case, strong latent heat fluxes over the ocean at \( 10^\circ \) increase the subcloud moist static energy which is advected onto the land. In the \( 8^\circ \) coastline case, the oceanic air which is advected across the coastline has lower moist static energy, and must travel further across the heated continent in order to reach the RCE \( h_b \).

6. Seasonal

The transient period prior to establishment of a steady circulation is nearly 200 days long for the continental perpetual summer cases - this is considerably greater than a seasonal timescale. It is not obvious that the nonlinear theory, which is based upon a steady-state circulation, is applicable to the highly transient monsoon. The delay in onset of jumping of the circulation questions the importance of the jumping behavior to the seasonal, transient monsoon. A series of experiments with seasonally varying land forcing are performed in order to investigate the pertinence of the steady state, perpetual summer results to the seasonal monsoon.

In order to retain simplicity, the ocean SST does not vary in time and is uniformly 302K at all latitudes. The coastline is located at \( \phi_e = 16^\circ \), and the land forcing is given by (6), with \( \phi_T = 8^\circ \). To represent seasonal variation in radiative forcing, \( THF_0 \) is varied sinusoidally in time from \( 80 \text{ W/m}^2 \) at winter solstice to the maximum summer value, with period of 365 days. The summer solstice magnitude of \( THF_0 \) is varied from \( 125 \text{ W/m}^2 \) to \( 150 \text{ W/m}^2 \), as in the perpetual summer cases. Due to the unrealistic choice of ocean surface temperatures, the results are not expected to be suitable for study of the dynamics
of monsoon onset. The model is first spun up in a winter solstice regime for 200 days, then the seasonal cycle is initiated, and the model is run for five annual periods. During the winter season, the cold landmass yields a direct meridional circulation cell with ascent over the northern hemisphere tropical ocean and subsidence over the cold continent. For all cases tested, the circulation and precipitation fields quickly adjusted to the seasonal cycle, and there was very strong interannual consistency. The last four years of the model run were averaged to create a mean annual progression.

For land forcing of $THF_0 \leq 130W/m^2$, the summer monsoon is weak, with a shallow capped meridional circulation over the subtropical continent. Little precipitation occurs over the continent during the summer season (Figure 10a). For stronger land forcing ($THF_0 \geq 140W/m^2$), large-scale ascent and deep convection form over the subtropical continent during the summer. The precipitation maximum over the boreal tropical ocean at first weakens during spring, then shifts poleward onto the continent and intensifies during the course of the summer (Figure 10b). At the beginning of fall, the continental precipitation maximum weakens abruptly, the rainfall peak then continues to migrate poleward during the winter until it dies out the following spring. During early summer, the meridional circulation is local (Figure 11a), with ascent over the subtropical continent and subsidence over the northern hemisphere tropical ocean. As the summer progresses, the circulation strengthens and broadens, with increased cross-equatorial flow. By the late summer, the circulation is quite broad, and jumping behavior occurs, as shown in Figure 11b.

The seasonal cases show that the steady state solutions are pertinent to the seasonal monsoon during late summer only. The mean summer net surface flux between June 1 and August 31 in the seasonal case is approximately $10W/m^2$ less than the solstice $THF_0$. In the seasonal experiment, the $THF_0 \leq 130W/m^2$ cases have a mean summer $THF_0 < 125W/m^2$ and do not result in a monsoon, which is similar to the perpetual summer results for $THF_0 < 125W/m^2$.

In the early and mid-summer, the steady state solution differs considerably from the
seasonal monsoon. The meridional circulation associated with the monsoon during early summer is localized in extent, confined to the boreal hemisphere. In mid June, the circulation becomes cross-equatorial and begins to fold over the contours of AAM in the upper troposphere. The jumping behavior which is a major feature of the steady state solutions is not observed until mid July in the seasonal cases.

7. Discussion

The monsoon is considered as a seasonal relocation of the ITCZ onto a subtropical landmass, and as such, the dynamics of the Hadley circulation are fundamental to the monsoon. The dynamics of the steady Hadley circulation have been explored in a series of analytic studies by Held and Hou (1980), Lindzen and Hou (1988), Plumb and Hou (1992), and Emanuel (1995). However, the simple analytic theory does not consider the effects of a subtropical landmass, which forces the atmosphere differently from an aquaplanet or simple applied forcing, and does not account for onset or transient behavior.

The axisymmetric theory is extended to predict the extent of the meridional circulation and the location of the monsoon. The location of the deep ascent branch of an AMC circulation is found to be strongly tied to the distribution of subcloud moist static energy. Given a local maximum of subcloud moist static energy, the poleward boundary of the circulation will be colocated with the maximum \( h_b \), and the large-scale ascent and precipitation will occur near and slightly equatorward of the maximum. Because the circulation itself interacts strongly with the subcloud moist static energy distribution, this theory is diagnostic rather than prognostic. However, the effect of various mechanisms upon the extent of the steady monsoon may be reduced to a determination of the impact upon the subcloud moist static energy. For example, orography can affect the subcloud moist static energy. Molnar and Emanuel (1999) have shown that the radiative-convective equilibrium surface air temperature decreases at a rate of approximately \( 2 \, K/km \) as the surface is raised, which is less steep of a decline than the moist adiabatic lapse rate. Assuming a moist adiabatic lapse rate,
the saturation entropy will be greater over an elevated surface than over a lower surface receiving the same incoming radiation, and the subcloud moist static energy will also be greater.

Two hallmarks of the nonlinear theory are examined to determine the nature of the flow induced by a local, subtropical forcing. One is the presence of threshold behavior, the other is the vanishing of absolute vorticity in the upper troposphere. In an aquaplanet setup with localized SST perturbation at 16N, the meridional circulation clearly exhibits threshold behavior, with a pronounced increase in the strength of the circulation for forcing above a certain magnitude. These strong circulations are nearly AMC, with upper tropospheric absolute vorticity close to zero. The threshold behavior in the aquaplanet case is possibly exaggerated by feedbacks between the latent heat flux at the surface and the circulation strength. When the SST perturbation is replaced with a subtropical continent, threshold behavior is not clearly seen when the land forcing strength is varied. However, the meridional circulations which develop for strong land forcing appear to be angular momentum conserving, with near zero absolute vorticity in the upper troposphere.

The lack of threshold behavior in the two dimensional cases with a subtropical continent in comparison to the prominence of the behavior in the aquaplanet situation is ascribed to a combination of factors. First, in the aquaplanet cases, there is a feedback between the circulation strength and the surface fluxes through the surface wind speed. This feedback tends to accentuate threshold behavior of the circulation strength in the aquaplanet case. In the continental cases, the net surface flux is prescribed, so that this feedback will not occur, and the circulation strength is not expected to show as strong an increase above the threshold as in the aquaplanet cases. A second, more subtle, factor is the poleward progression of the monsoon with increased forcing in the continental cases. A series of aquaplanet cases with varied location of the SST perturbation has shown that the circulation weakens as it extends further poleward. As the monsoon moves poleward in the continental cases, broadening the circulation, the circulation strength is expected to weaken somewhat, obscuring threshold
behavior.

How well do the modeled circulations conform to the requirement of colocation of the circulation boundary with the maximum of subcloud moist static energy? The cases with subtropical continent uphold the theory quite well: the poleward boundary of the meridional circulation is colocated with the maximum subcloud moist static energy, and the monsoon precipitation occurs slightly equatorward of this maximum. The theory holds reasonably well for the aquaplanet cases, but there are a few examples in which the boundary of the circulation occurs poleward of the subcloud moist static energy maximum, and the ascending branch of the circulation has westerly shear with height. In these cases, the numerical filters break down angular momentum conservation in the ascent branch of the circulation, and the theory does not apply.

According to the theory laid out by Plumb and Hou (1992), the forcing strength required to induce an AMC circulation is dependent upon both the shape and location of the forcing: the closer to the equator, the less forcing is required. This is borne out by the modeled cases with the coastline located at 8N. In these cases, a global AMC circulation occurs at a weaker forcing level than in the cases with coastline at 16N, and the circulation is more vigorous. When the coastline is moved to 24N, an AMC circulation may still result for sufficiently strong forcing, but the cell is local and not global, following the theory of Schneider (1983).

The nonlinear theory concerns the steady state circulation, but the real monsoon is a transient, seasonal phenomenon. The timescales needed to reach a steady state in the model are quite long, especially in the two dimensional cases. In the experiments with time-varying land forcing, the transient monsoon circulation only bore a resemblance to the steady state during the late summer period. However, the ocean SST distribution used in these cases was far from realistic, which impacts the meridional circulation. The initial adjustment period occurs while the circulation folds over contours of constant angular momentum in the upper troposphere. In the real world, an cross-equatorial Hadley circulation forced by the ocean gradient exists prior to the onset of the monsoon, so that the momentum field is already
significantly rearranged when the monsoon initiates. Fang and Tung (1999) found that the transient circulation was similar in strength to the steady circulation for off-equatorial forcing, which supports the idea that the steady state dynamics may be applicable to the seasonal case.

The idealized physics used in this study neglects many important processes which may have substantial impacts upon the monsoon. The version of the MITGCM used here does not support orography, which may strongly influence the monsoon, as previously discussed. The model has only the simplest representation of boundary layer physics, consisting of a momentum mixed layer of fixed depth and dry adiabatic adjustment of variable depth. The land surface hydrology is very primitive, and there is no allowance for vegetation or other surface biosphere components. Given the importance of boundary layer thermodynamics to the large scale monsoon which has been emphasized in this study, the simplicity of the boundary layer representation in the model is a serious limitation. One important process which was omitted from the model is radiation. Although the use of Newtonian cooling makes the dynamics much more straightforward, the prescribed net radiative flux used to calculate the surface energy balance is not very realistic. The model neglects feedbacks from clouds, the radiative influence of water vapor, longwave feedbacks from the surface temperature, albedo and vegetative effects on radiation, all of which may impact the monsoon. A diurnal cycle is also not included in the model, which may impact the steady state.

Future work is suggested to investigate how different physical processes impact the subcloud moist static energy and the large scale monsoon circulation. Important processes which were neglected from the idealized model used here include radiative feedbacks, orographic effects, and boundary layer physics. The theory linking the monsoon location to the subcloud moist static energy maximum may also be able to help explain transient monsoon features, such as onset and break monsoon.
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List of Figures

1 Meridional SST profile (K) used in aquaplanet cases with localized subtropical SST perturbation of magnitude $\Delta T$ at $\phi_0$. .......................... 30

2 Steady-state results as a function of subtropical SST forcing for aquaplanet case. Top, absolute global minimum circulation streamfunction strength, $kg\, s^{-1}$. Bottom, minimum 150 hPa absolute vorticity between 6°N and 64°N. 31

3 Steady-state streamfunction for aquaplanet case, 100 day time mean, SST perturbation located at $\phi_0 = 16^\circ N$. Solid contours denote counter-clockwise flow, dashed contours indicate clockwise flow. Note that contour interval is 4 times greater for the figure on the right. Left, subthreshold result for $\Delta T = 1.0K$, contour interval $2.5E9kg\, s^{-1}$. Right, supercritical result for $\Delta T = 2.0K$, contour interval $1E10kg\, s^{-1}$. .......................... 32

4 Absolute global minimum steady-state streamfunction for aquaplanet case, 100 day time mean, SST perturbation at northern hemisphere latitude $\phi_0$. 33

5 Steady-state streamfunction for continental case with uniform warm ocean, 100 day time mean, coastline located at $\phi_L = 16^\circ N$. Solid contours denote counter-clockwise flow, dashed contours indicate clockwise flow. Left, subcritical result for $THF_0 = 130W\, m^{-2}$, contour interval $5E9kg\, s^{-1}$. Right, supercritical result for $THF_0 = 140W\, m^{-2}$, contour interval $5E9kg\, s^{-1}$. 34

6 Steady-state results as a function of land forcing strength. Continent at $\phi_0 = 16^\circ N$ with uniformly warm ocean, dotted lines with squares; continent at $\phi_0 = 16^\circ N$ with “summer” SST, dot-dash line with asterisks; continent at $\phi_0 = 8^\circ N$ with “summer” SST, solid lines with circles; continent at $\phi_0 = 24^\circ N$ with “summer” SST, dotted lines with triangles. Top, absolute global minimum circulation streamfunction strength, $kg\, s^{-1}$. Bottom, minimum 150 hPa absolute vorticity between 6°N and 64°N. 35
Location of maximum monsoon precipitation for steady monsoon as a function of land surface forcing strength ($THF_0$). Case with coastline at $\phi_L = 16^\circ N$ and uniform warm SST. ................................. 36

Steady-state streamfunction for cases with uniform warm ocean, 100 day time mean for cases with summer-like SST distribution. Solid contours denote counter-clockwise flow, dashed contours indicate clockwise flow, contour interval $1E10kg s^{-1}$. Top left, aquaplanet case with SST maximum at $8^\circ N$; Top right, continental case with coastline at $\phi_L = 16^\circ N$ and $THF_0 = 140W m^{-2}$; bottom left, continental case with coastline at $\phi_L = 8^\circ N$ and $THF_0 = 140W m^{-2}$; bottom right, continental case with coastline at $\phi_L = 24^\circ N$ and $THF_0 = 140W m^{-2}$. ................................. 37

Schematic diagram of subcloud moist static energy. Dashed line shows radiative convective equilibrium $h_b$, solid line shows $h_b$ in the presence of a large-scale circulation. ................................................. 38

Hovmöller diagram of precipitation for seasonal cases, contour interval 2.0 $mm day^{-1}$. Left, maximum summer land forcing $THF_0 = 130W m^{-2}$; right, maximum summer land forcing $THF_0 = 150W m^{-2}$. Day zero occurs at winter solstice, annual mean data over four years of model integration. .... 39

Summer circulation for seasonal case with $THF = 0 = 150W m^{-2}$, contour interval $5E9kg s^{-1}$. Left, early summer circulation (one week before summer solstice); right, late summer circulation (late August). ....................... 40

Steady-state fields for continental case with $THF_0 = 150W m^{-2}$, coastline at $\phi_L = 16^\circ N$, with uniform warm ocean. Top, streamfunction, contour interval $5E9kg s^{-1}$. Center top, zonal wind, contour interval $10m s^{-1}$. Center bottom, precipitation, $mm day^{-1}$. Bottom, 1000 mb moist static energy, $10^4 J$. .... 41

29
Figure 1: Meridional SST profile (K) used in aquaplanet cases with localized subtropical SST perturbation of magnitude $\Delta T$ at $\phi_0$. 
Figure 2: Steady-state results as a function of subtropical SST forcing for aquaplanet case. Top, absolute global minimum circulation streamfunction strength, kg s\(^{-1}\). Bottom, minimum 150 hPa absolute vorticity between 6°N and 64°N.
Figure 3: Steady-state streamfunction for aquaplanet case, 100 day time mean, SST perturbation located at $\phi_0 = 16^\circ N$. Solid contours denote counter-clockwise flow, dashed contours indicate clockwise flow. Note that contour interval is 4 times greater for the figure on the right. Left, subthreshold result for $\Delta T = 1.0K$, contour interval $2.5E9 kg s^{-1}$. Right, supercritical result for $\Delta T = 2.0K$, contour interval $1E10 kg s^{-1}$. 
Figure 4: Absolute global minimum steady-state streamfunction for aquaplanet case, 100 day time mean, SST perturbation at northern hemisphere latitude $\phi_0$. 
Figure 5: Steady-state streamfunction for continental case with uniform warm ocean, 100 day time mean, coastline located at $\phi_L = 16^\circ N$. Solid contours denote counter-clockwise flow, dashed contours indicate clockwise flow. Left, subcritical result for $THF_0 = 130 W m^{-2}$, contour interval $5E9 kg s^{-1}$. Right, supercritical result for $THF_0 = 140 W m^{-2}$, contour interval $5E9 kg s^{-1}$.
Figure 6: Steady-state results as a function of land forcing strength. Continent at $\phi_0 = 16^\circ N$ with uniformly warm ocean, dotted lines with squares; continent at $\phi_0 = 16^\circ N$ with “summer” SST, dot-dash line with asterisks; continent at $\phi_0 = 8^\circ N$ with “summer” SST, solid lines with circles; continent at $\phi_0 = 24^\circ N$ with “summer” SST, dotted lines with triangles. Top, absolute global minimum circulation streamfunction strength, $kg \, s^{-1}$. Bottom, minimum 150 hPa absolute vorticity between $6^\circ N$ and $64^\circ N$. 
Figure 7: Location of maximum monsoon precipitation for steady monsoon as a function of land surface forcing strength ($T HF_0$). Case with coastline at $\phi_L = 16^\circ N$ and uniform warm SST.
Figure 8: Steady-state streamfunction for cases with uniform warm ocean, 100 day time mean for cases with summer-like SST distribution. Solid contours denote counter-clockwise flow, dashed contours indicate clockwise flow, contour interval $1E10\text{kg s}^{-1}$. Top left, aquaplanet case with SST maximum at $8^\circ N$; Top right, continental case with coastline at $\phi_L = 16^\circ N$ and $THF_0 = 140W\text{ m}^{-2}$; bottom left, continental case with coastline at $\phi_L = 8^\circ N$ and $THF_0 = 140W\text{ m}^{-2}$; bottom right, continental case with coastline at $\phi_L = 24^\circ N$ and $THF_0 = 140W\text{ m}^{-2}$. 
Figure 9: Schematic diagram of subcloud moist static energy. Dashed line shows radiative convective equilibrium $h_b$, solid line shows $h_b$ in the presence of a large-scale circulation.
Figure 10: Hovmoeller diagram of precipitation for seasonal cases, contour interval 2.0 mm day$^{-1}$. Left, maximum summer land forcing $THF_0 = 130W m^{-2}$; right, maximum summer land forcing $THF_0 = 150W m^{-2}$. Day zero occurs at winter solstice, annual mean data over four years of model integration.
Figure 11: Summer circulation for seasonal case with $THF - 0 = 150 W m^{-2}$, contour interval $5 E 9 kg s^{-1}$. Left, early summer circulation (one week before summer solstice); right, late summer circulation (late August).
Figure 12: Steady-state fields for continental case with $THF_0 = 150 W m^{-2}$, coastline at $\phi_L = 16^\circ N$, with uniform warm ocean. Top, streamfunction, contour interval $5 E 9 kg s^{-1}$. Center top, zonal wind, contour interval $10 m s^{-1}$. Center bottom, precipitation, $mm day^{-1}$. Bottom, 1000 mb moist static energy, $10^4 J$. 