Abrupt Onset and Slow Seasonal Evolution of
Summer Monsoon in an Idealized GCM Simulation

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Abstract

The world's greatest monsoons and deserts co-exist at the same latitudes on the subtropical Afro-Eurasian Continent. To investigate the mechanisms for these diverse subtropical climates, we conduct atmospheric general circulation model experiments under an idealized land-sea distribution featuring an Asian continent with a straight south coast at 17°N. Sea surface temperature and land surface parameters are all set zonally uniform to avoid a priori longitudinal preference for rainfall. The model with flat land surface produces a realistic zonal distribution of precipitation, with arid (wet) conditions over the western (eastern) part of subtropical Afro-Eurasia in association with a gigantic high pressure system over the ocean surface. The key to maintaining the western deserts is the slow precipitation-soil moisture interaction, which keeps the northward migration of the monsoon rainbelt to lag behind the sun. This is demonstrated by conducting a perpetual summer integration, where the Sahara and Arabia are soaked in heavy rain.

The model monsoon starts abruptly in late June, three months after the spring equinox when a northward temperature gradient is established near the ground on the south coast. The onset is associated with explosive growth of a westward-traveling moist baroclinic wave. Further analysis suggests a Geostrophic Monsoon model. The geostrophy resists and delays the formation of a thermally direct circulation, until baroclinic instability triggers the rapid onset.

1. Introduction

The southeast and south Asian monsoons dominate the summer climate over much of the subtropical Asia, the northwest Pacific and the North Indian Oceans. The monsoon rainfall first starts over Indochina in May and then moves northward to cover the Indian subcontinent. On large scales, the intense Somali Jet and low-level westerlies blow over the North Indian Ocean after the onset of the Indian monsoon. At the height of the summer monsoon, tropospheric temperature reverses its meridional gradient and increases poleward between the equator and 30°N, and a gigantic anticyclone dominates the upper level flow field in the Northern Hemisphere. See Li and Yanai (1996), Barry and Chorley (1992) and Lau and Yang (1996) for more description of the monsoon and its seasonal march.

In contrast to the gradual seasonal increase in solar radiation, the Indian Monsoon is known for its abrupt onset around mid-June, a dramatic event whose mechanisms are poorly understood. Figure 1 shows time series of surface temperature and outgoing longwave radiation (OLR) over India from April 1 to October 1, 1993. Before the monsoon starts, ground is dry and intense insolation raises surface temperatures over India to as high as 37°C in daily average. This hot premonsoon period starts in late April and lasts for one and a half months. Then in mid-June, deep convection and precipitation take place as indicated by a sharp decrease in OLR. With sudden arrival of monsoon rains, surface temperatures drop sharply below 30°C in just two days, due to both cloud shielding and evaporative cooling.

The summer monsoon is often described as a planetary-scale sea breeze caused by a temperature difference between hot land and cool ocean surface. This sea-breeze paradigm, however, does not explain either the abruptness or the time delay of the onset, two hallmarks of the Indian monsoon. While low-level temperature over land exceeds that over the Indian Ocean in as early as late April, the Indian
monsoon does not start until one and a half months later. This delayed onset of subtropical monsoon seems due to the effects of Earth's rotation. In response to a given thermal forcing, a rotating atmosphere does not necessarily produce a meridional circulation as depicted by the sea breeze paradigm. Instead it can adjust its vertical shear so as to be in the thermal wind balance with meridional temperature gradient (Held and Hou, 1980). In an axisymmetric model, Plumb and Hou (1992) show that such a thermal wind adjustment is possible for an off-equatorial heating, and that an intense meridional circulation occurs only when the subtropical heating rate exceeds a certain threshold (Emanuel, 1995). Krishnakumar and Lau (1998) propose that the inertial instability near the equator may trigger this transition into a thermal direct circulation. In reality, heating associated with deep convection cannot be prescribed but has to be determined from the interaction with a flow field. In some latitude-height two-dimensional (2D) models that allow this interaction, the summer monsoon starts soon after the spring equinox with little time delay (Webster and Chou, 1980).

While simple models tend to emphasize a 2D picture, the real monsoons have an extra dimension in the east–west direction. Westward or northwestward traveling monsoon depressions are the well-known variability of the monsoon on subseasonal time scales. These low pressure systems form in the eastern Bay of Bengal at a frequency of twice a month from June to September. Saha et al. (1981) find that 45 of the 52 depressions in a ten-year period, their formation can be traced further eastward, to southeast Asian and the South China Sea. Hastenrath (1991) suggests that a major portion of monsoon precipitation over India may be attributable to these depressions. These westward propagating disturbances are the source of the bi-weekly oscillation (Krishnamurti and Bhalme, 1976) over the Indian Subcontinent, as clearly visible in our post-onset OLR time series (Fig. 1). An analysis of the 1993 summer OLR indicates that the bi-weekly oscillations over India is associated with westward-moving disturbances originating from the western Pacific (not shown), consistent with Saha et al. (1981). It is unclear how these westward traveling waves might fit into the 2D picture, and whether they are just some non-essential details, or might impact significantly on the onset and evolution of the monsoon.

World's greatest monsoons, and deserts (Sahara and Arabia), coexist on the subtropical Afro-Eurasian continents, further contradicting simplistic 2D models of monsoon. Based on a model forced by prescribed heating, Rodwell and Hoskins, (1996) suggest that baroclinic Rossby waves excited by the South Asian monsoon induce downrake and maintain the deserts to the west. An interesting deviation from the traditional 2D explanation of subtropical deserts, this desert-causing mechanism needs to be tested in a model that allows interactions of heating and flow fields. As an elevated sensible heat source in summer (Yeh et al., 1957; Hahn and Mababe, 1975; Yanai et al., 1992), the Tibetan Plateau can cause zonal rainfall variations on the subtropical Afro-Eurasian continents. The possibility of other complementary causes of this eastern monsoon–western desert configuration has been left largely unexplored. Here we will suggest a slow interaction of monsoon and ground hydrology as such a possible cause.

As reviewed above, our understanding of monsoon is largely based on two extreme types of models. At one extreme are latitude-height 2D models with specified heating. At the other extreme are full-fledged general circulation models (GCMs) run under realistic boundary conditions with an aim to simulate every detail of the real atmosphere. The present study attempts to establish a middle point that bridges the above two classes of models. Our approach is to run a GCM with full physics albeit under a highly idealized land-sea configuration (Fig. 1) and examine a simple monsoon induced by land-sea difference only. We ask the following questions: With land-sea difference as the sole zonal varying boundary condition, will the eastern monsoon–western desert configuration appear, and why? Does this simple monsoon start smoothly or abruptly, with or without time delay behind the equinox? What controls the monsoon's subsequent development? In particular, how far can the summer
rainbelt penetrate inland, and what controls this penetration distance? The most important results are that baroclinic instability triggers the abrupt onset and that the slow adjustment of soil moisture field limits how far the monsoon can penetrate inland.

Simplified surface boundary conditions can sometimes offer unique insights into the essence of the phenomenon of interest. A successful example is the so-called aqua-planet experiments (Hayashi and Sumi, 1986; Numaguti and Hayashi, 1991; Hess et al., 1991; Xie et al., 1993), which revealed the self-organizing nature of tropical moist convection on the planetary scale. The particular monsoon experiment here is designed to free us from considering complicated features such as coastal lines, land topography and zonal variations in SST, and land surface parameters. As such, our main purpose here is not to mimic every detail of the real monsoon, but rather to reveal a first order picture of monsoon and its cause. The results provide a prototype based on which more experiments can be conducted. By adding more realistic surface features, we can recover the real monsoon with full complexity and delineate the effects of each added feature. Such an example is provided toward the end of the paper, where a "Tibetan" plateau is introduced and its effects examined.

Whereas recent studies have shifted the focus to monsoon's interannual variability and its causes such as continental snow cover variations (Hahn and Shukla, 1976; Barnett et al., 1989; Yasunari et al., 1991; Meehl, 1994), and equatorial Pacific SST anomalies (Palmer et al., 1992; Nigam, 1994; Shen et al., 1998; Yang and Lau, 1998), the inter-GCM variability in seasonal-mean rainfall distribution remains uncomfortably high, comparable to or even larger than interannual variability within individual models (Lau et al., 1996). Revisits to the fundamentals can improve our understanding of monsoon dynamics, and thus help reduce the inter-GCM variability.

In the rest of the paper, Section 2 describes the GCM and its boundary conditions. Section 3 contrasts the runs with insolation varying seasonally and fixed at its summer value, thus illustrating the effect of ground hydrology. We will focus on what controls the northward progression of monsoon rain. Section 4 investigates the mechanism for the sudden monsoon onset in the model. Section 5 is a summary, and proposes a new conceptual model of summer monsoon.

2. Model

The GCM is developed jointly by University of Tokyo's Center for Climate System Research (CCSR) and National Institute for Environmental Studies (NIES) in Japan. It solves primitive equations of motion in spherical spectrum form. Here we employ a version with triangle truncation at zonal wavenumber 21 (T21), and 20 sigma levels in vertical. The model employs a modified Arakawa-Schubert (1974) moist convection scheme. Cloud water is predicted (Le Treut and Li, 1992) and radiation is calculated according to Nakajima and Tanaka (1986). Readers are referred to Numaguti et al. (1997) and Numaguti (1999) for details of model physics and its performance. In particular, the model successfully simulates both the climatology and interannual variability of the Asian summer monsoons (Shen et al., 1998).

Here we prescribe an idealized land-sea configuration (Fig. 2). The Antarctic is circular with the north coast at 69°S. The Afro-Eurasian continent occupies one half of a longitudinal cycle between 90°E to 90°W. The model Africa has a bulge on the west coast and its east coast lies along 140°E. The south coast of Asia is straight oriented in the east-
west direction and the first land grid point is placed
at 19°N. Unless stated otherwise, land surface is flat
at the sea level. In only one case, a gentle Gaussian
topography with a maximum height of 4000 m, is
placed within 170°E–60°W and 19°N–50°N.
Ground hydrology is represented by a bucket
model 15 cm deep (Manabe et al., 1965). Predicted
from precipitation, evaporation and runoff,
soil moisture content in the bucket determines the
ratio of actual to potential evaporation. Because of
its simplicity, this bucket has been widely used for
a long time in GCM modeling. We feel it adequate as
a first approximation of hydrological processes, com-
patible to our idealized experimental design. Build-
ing on this, future work can take advantage of recent
progresses in land surface modeling to include more
sophisticated physics and plant physiology. Land
surface albedo is prescribed zonally uniform at the
zonal average of observed monthly climatology over
land surface below 2000 m. It will be interesting to
incorporate in the future a vegetation model with in-
teractive surface albedo and water storage capacity,
which might provide a positive feedback in desert
formation (Charney, 1975).
Sea surface temperature field is also zonally un-
iform and computed by zonally averaging observed
monthly climatology. In the Tropics equatorward
of 30°, only SSTs in the Indian and western Pa-
cific Oceans (40E–180) are used in the zonal average
to exclude equatorial cold tongues in the Atlantic
and eastern Pacific. Figure 2 displays the latitude-
time section of prescribed SST field. The warm
band with SST higher than 28°C migrates north-
southward smoothly following the seasonal march of
the sun. Monthly maximum SSTs over the Tropics
exceed 29°C in March through May and then
decrease with time, a manifestation of cooling in the
Indian Ocean due to the strengthening of surface
monsoonal winds. The SST in the real world varies
interactively with the atmosphere. The South Asian
monsoon affects the North Indian Ocean SST while
the ENSO gives rise to large variability in equatorial
SST. Using the same GCM under realistic boundary
conditions, Shen et al. (1998) show that the onset
date of the Indian monsoon varies with equatorial
Pacific SSTs. Zonally uniform SST distribution, as
is prescribed in our idealized experiments, appears
on the equator in the boreal springs of 1983 and
1998, years of strong El Niño.

The realistic version of the model is first spun up
from isothermal initial conditions for ten years and
the zonal means of the last year simulation is used to
initialize the idealized version. The model then
is run for multiple years, and results excluding the
first spin-up year will be discussed below.

3. Slow seasonal evolution
In this section we describe the seasonal develop-
ment of monsoon in the model, in particular how the
flow field and ground hydrology interact to move the
monsoon rain inland and differentiate wet and arid
regions on the continent. The next section will shift
attention to subseasonal aspects of monsoon, most
notably its onset.
3.1 Seasonal run
Figure 3 show latitude-time sections of monthly
mean precipitation based on two-year long, three
daily model outputs. With land in both hemispheres
over Africa, monthly-mean rainfall distribution has
a single peak all the time, which migrates north-
southward following the sun. This rainfall peak
reaches its northernmost position at 14°N in August
when precipitation covers a broad latitudinal band
from the equator to 20°N. In contrast to this grad-
ual northward migration over Africa, precipitation
sets into the Northern Hemisphere rather abruptly
in June over the model Indian Ocean. In boreal
summer, there are two precipitation maxima, one
on the equator and the other on the south coast of
the continent. The monsoon rain band in the model
has a broad maximum spanning both the oceanic
and land grid points. This rainbelt penetrates as far
north as 30°N. South of the mid-latitude rainbelt loc-
ated around 50°N, summer is the only season for
the continent to receive rainfall, which is under the
control of the downdraft of the Hadley cell in win-
ter. Thus between 30°N and 45°N lie arid regions
that receive little rain throughout the year and may
correspond to Central Asia. Without mediterranean
seas and mountains, the mid-latitude deserts extend
from the west to the east coast in the model. The
southern edge of this desert band in the model is
determined by how far the monsoon rain can penetrate
inland in summer.

It is interesting to note that briefly in March
and April when the SST distribution has a broad
maximum near the equator, the model develops
twin intertropical convergence zones (ITCZs), one
in each hemisphere. The ITCZ configuration on a
zonally uniform aqua-planet with a broad equa-
torial SST maximum is sensitive to cumulus pa-
rameterization schemes (Numaguti and Hayashi,
1991; Hess et al., 1991), and its mechanism is
not well understood. The GCM has a bias for
equatorial precipitation. In boreal summer the
equatorial rainbelt is much too strong compared
to the satellite measurements. The tropical in-
traseasonal oscillation is not well simulated either,
with equatorial precipitation showing a spurious
tendency for westward propagation throughout the
year (not shown). We will not discuss the equa-
torial rainfall further in the rest of the paper.
Figure 4 shows summer precipitation and 850 hPa wind velocity fields. The monsoon rainbelt tilts northeastward, covering the eastern half of the subtropical continent but leaving the western half without rain north of 20°N. It is remarkable that in the absence of mountains and zonal variations in land surface albedo, the model still reproduces the contrast between the wet monsoon climate in east, and deserts in the west. This northeastward tilt can be explained by flows in the lower atmosphere. Because of the land-sea heat contrast, a huge anti-cyclone forms over the subtropical ocean. The winds associated with this oceanic anti-cyclone brings moist air inland in the eastern continent while pushing down the rainbelt in the west by transporting dry air from mid-latitudes southward. The oceanic anti-cyclone penetrates deep into the western continent as observed over South Europe and North Africa. This contrast in summer rainfall between the eastern and western subtropical continent remains in a model run where Africa is removed and the south coast line runs straight along 19°N from 90°E to 90°W.

In boreal summer, a local rainfall maximum is observed over the equatorial Indian Ocean, separate from intense monsoon precipitation to the north. The model reproduces this rainfall configuration except that the equatorial rainbelt is much too strong. In the model, the latent heating associated with
equatorial rainfall has a shallow and bottom-heavy vertical structure that peaks just above the planetary boundary, whereas the heating by monsoon rainfall has a deep vertical structure peaking at 500 hPa (Fig. 5). In addition to these tropical rainbands, there are a pair of mid-latitude rainbelts associated with synoptic disturbances in the westerly jets in both hemispheres.

### 3.2 Perpetual runs

The heat content of land surface is so small compared to that of the ocean that the atmosphere is often assumed to adjust to a quasi-steady state within a month in response to changes in external forcing. This quasi-steady approximation leads to the development of simple slave atmospheric models (Matsumoto, 1966; Gill, 1980), which are instrumental in rapid progress in coupled ocean-atmosphere modeling of the mean state (Xie 1996) and variability of the tropical climate (Neelin et al., 1998; Xie et al., 1999). Perpetual GCM runs with fixed insolation are a widely used technique in planetary wave and other general circulation studies. Previous perpetual summer runs (Pitcher et al., 1983; Kutzbach et al., 1989) confirm that the atmosphere adjusts to a statistically steady state rather fast, provided that soil moisture is held constant in time. Here we systematically test this quasi-steadiness assumption by conducting twelve (12) perpetual runs with insolation and SST fixed at their values on the first day of each calendar month. Soil moisture field is allowed to interact with the atmosphere to see how land hydrology affects the model adjustment. In each run, the model is integrated for 12 months and the average of the last four months will be shown. So each perpetual run can be viewed as the model’s steady state response to an insolation distribution and deviations from it in the seasonal run will be attributed to some slow, unsteady processes in the model.

Perpetual runs reproduce the precipitation distribution of the seasonal run in most months except for boreal summer (Fig. 6). The summer climate in the perpetual June 1, July 1 and August 1 runs is drastically different from that in seasonal runs. Monsoon rains penetrate deep inland to as far north as 40°N in both the African and Indian Ocean sectors. Figure 7a displays the precipitation distribution in the perpetual June 1 run. While tropical precipitation zones over oceanic regions remain largely unchanged compared to the seasonal run, the great deserts of the Sahara and Arabian peninsula are soaked in heavy rains. Adding a “Tibetan Plateau” into the model does not help the matter, only marginally modifying the precipitation distribution (not shown).

### 3.3 Roles of ground hydrology

The interactive ground hydrology is what drives the summer climate away from the quasi-steadiness. Soil moisture adjustment is a slow process, with longer time scales than the thermal and dynamic wave adjustments (Yeh et al., 1984). Because the downdraft of the winter Hadley cell empties the buckets of the model over the subtropical continent, it takes one month to re-fill at a typical precipitation minus evaporation rate of 5 mm/day. To illustrate the role of this slow bucket-filling process, we conduct a pair of parallel experiments: one usual seasonal run, and one perpetual June 1 run initialized with the June 1 fields of the former.

Figure 8b shows the latitude-time section of 140–200°E average rainfall along with soil moisture content in the perpetual run. On June 1, the continent is dry and precipitation is localized to a latitudinal
band slightly south of the equator. This precipitation configuration with a single equatorial rainbelt persists for a month, and then suddenly a Northern Hemisphere (NH) rainbelt emerges off the coast. This NH rainbelt then gradually progresses northward. In September, it centers on 25°N and brings rainfall as far inland as 30°N.

To see how the monsoon rainbelt advances inland in the perpetual run, we average the model fields in a coordinate moving northward with the rainbelt at a speed of 12 km/day. Both local evaporation and remote moisture supply via converging wind contribute to the monsoon rainfall. At the peak of the composite precipitation distribution, about half of the rainfall is supplied by local evaporation, whose relative contribution increases further inland (Fig. 9). Specific humidity reaches maximum at the precipitation peak and decreases rapidly northward. Within the rainbelt, clouds shield of insolation and evaporation keep surface temperatures from rising. North of the rainbelt, clear sky allows the intense solar radiation to heat the dry ground, and temperatures are 15°C higher than within the rainbelt. The rapid northward increase of surface temperature shifts the moist static stability peak 5° north of the existing rainbelt. This phase difference between the moist static stability and precipitation causes the rainbelt to advance northward. This mechanism for northward precipitation propagation is unique to the continental hydrological cycle. Over the ocean, specific humidity and moist static stability are in phase with precipitation so SST can be regarded as fixed on the atmospheric time scales. A similar mechanism is previously proposed by Webster (1983). Although he emphasizes its role in causing sub seasonal variability, a slow seasonal march of rainfall toward the north is visible in his model, presumably by the same mechanism as in the present GCM.

The northward progression of the monsoon rain is accompanied by a front of soil moisture content, located north of the precipitation center and separating northern dry and southern wet regions. Wet ground seems a necessary condition for deep moist convection over land where local evaporation constitutes a major portion of precipitation and is regulated by ground wetness. This is demonstrated by an additional perpetual June 1 experiment, where soil moisture field is kept at its June 1 values. The NH precipitation band becomes much weaker on the continent because dry ground prevents it from moving far inland (see also Shakla and Mintz, 1982). Thus, a prescribed soil moisture field pre-determines where precipitation occurs over land. In reality, soil moisture and precipitation are an interacting pair so neither can be predicted a priori. Inland drizzles moisten soil while wet ground in turn supplies moisture to sustain the deep convection and rainfall. This interaction of rainfall and soil moisture is a slow process, limiting the speed of the northward

1 Without local supply by evaporation, moisture advected inland diffuses rapidly both vertically and horizontally within the deepening mixed layer. As a consequence, the lower atmosphere cannot attain a moisture content high enough for deep convection. This hypothesis needs to be tested in the future. Experiments with a high-resolution regional model indicate that the development of squall lines is sensitive to vertical structure of the deep mixed layer over dry land surface (Takemi and Satomura, 1999).
Fig. 7. Precipitation distributions [light (dark) shade > 3 (6) mm/day] in the perpetual June 1 runs (a) with interactive and (b) fixed soil moisture field.

Fig. 8. Latitude-time sections of 140-200E average precipitation [medium (dark) shade > 4 (8) mm/day] and soil moisture (contours: 0.05 & 0.1 m/m) in (a) seasonal run; and perpetual June 1 insolation runs with (b) interactive and (c) fixed June 1 soil moisture fields. Runs in (a) and (b) have the same initial conditions. Light shade denotes precipitation rate greater than 2 mm/day in (c).
penetration of the monsoon rainbelt.

The rainfall distribution in Fig. 7a can be considered as the model steady state in response to the summer insolation. Because of the slow time scale of soil moisture adjustment, it takes the model several months to approach this steady state. Before this could happen, the sun starts to move southward in the seasonal run and the decreased solar radiation pulls the NH rainbelt toward the south in September, terminating the boreal summer monsoon (Fig. 8a). Results from perpetual summer runs demonstrate that under steady summer insolation, the interaction of the atmosphere and ground hydrology has a potential to bring rains to the whole subtropical continents including the now arid Saharan and Arabian deserts. This potential is not realized because the northward migration of monsoon rainbelt is too slow and lags far behind the sun. The response of the climate system to summer insolation is intrinsically unsteady, and it is because of this unsteady nature that we observe the present distribution of deserts and monsoons on Earth.

Perpetual GCM experiments with fixed soil moisture are often conducted in paleoclimate research. In response to increased seasonal insolation, the increase in African summer rainfall in these experiments is modest, and tends to be limited to the present wet regions over and south of the Sahel, leaving precipitation over most of the Saharan largely unchanged (Kutzbach et al., 1996). Even with interactive soil moisture, current generations of atmospheric GCMs is known to severely under-estimate Saharan wet events both in intensity and meridional extent compared to geological evidence. The interaction of precipitation and soil moisture as revealed in the above analysis may be useful for improving paleo-simulation. For example, an increase in solistic insolation as in 12,000 year before present could increase precipitation over Africa, accelerating the northward migration of the rainbelt. An increase in wet area in turn increases moisture supply, leading to enhanced precipitation. This positive feedback between precipitation and northern reach of the monsoon could be important in causing the large wet anomalies over the Sahara. This hypothesis needs to be tested with improved ground hydrological models.

4. Abrupt onset

The appearance of the NH rainband is an abrupt event in the model (Fig. 8a). The longitude-time section of precipitation on the south coast of the model Asia (Fig. 10a) indicates that the monsoon’s onset, at a given longitude, is associated with the arrival of a westward moving rainband, which is first seen on the east coast around mid-June, and takes about one month to reach Africa. After the onset, the monsoon rainfall is punctuated by similar westward propagating waves. The wave period at a given location is about two weeks.

To highlight these bi-weekly waves, we remove the seasonal trend defined as one-month running means, and show in Fig. 10b the high-passed precipitation rate and relative vorticity at 850 hPa. The precipitation waves are coupled with vorticity waves, with the latter leading the former’s in phase. Vorticity anomalies are generally better organized than precipitation and can be traced around the globe. Even prior to the onset disturbance that brings out heavy precipitation to the continent, vorticity anomalies
are organized into packets of westward propagating waves. The two wave packets start to develop in mid-May and are accompanied by drizzles. Before that, vorticity disturbances are less well organized and show somewhat eastward phase propagation, probably steered by the subtropical westerly jet. In the subtropical Southern Hemisphere, the disturbances also show a weak tendency for eastward phase propagation throughout the year. All this suggests that the strong westward-moving bi-weekly disturbances are unique to the northern summer monsoon.

4.1 Pre-monsoon mean state

Figure 11 shows the pre-monsoon environment in which the first westward-moving wave packets develop. The subtropical continent receives little rain during the winter, so its ground is dry. Increased summer solar radiation heats ground rather efficiently, and causes intense dry convection reaching as high as 550 hPa. As a result, a northward temperature gradient is maintained in the lower atmosphere between the cool oceanic, and heated land regions, whereas temperature in the upper atmosphere remains decreasing poleward. In thermal wind balance with the poleward increase in lower-level temperature, is a tropical easterly jet centered on 15°N and 700 hPa. There is a deep westerly jet increasing its speed upward in mid-latitudes where the remaining winter snow reflects solar radiation and maintains a negative meridional temperature gradient near the ground.

The dry convection over the continent drives a shallow overturning cell while downdraft dominates the middle and upper troposphere (Fig. 11c). This pre-monsoon shallow cell may lead to a spurious early onset if a model has insufficient vertical resolutions and/or employs a parameterization that ties deep convection too strongly to low-level moisture convergence. This may explain the early monsoon onset in Webster and Chou’s (1980) two-level
model, where the low-level convergence occurs over land soon after the spring equinox in response to increased solar heating. Sufficient vertical resolutions to resolve the non-precipitating shallow overturning cell are thus crucial to a successful simulation of the delayed monsoon onset. In our GCM, the shallow overturning cell is capped by downward motion in the upper troposphere, where the temperature field still maintains a southward gradient.

Figure 11b shows the latitude-pressure section of Ertel’s potential vorticity (PV). Weak temperature stratification over southern continent leads to the formation of a potential vorticity minimum layer that extends from the ground to 400 hPa. With easterly shear/poleward temperature gradient near the ground, the negative PV gradient within 5°N–25°N indicates that this premonsoon mid-level easterly jet may be baroclinically unstable (Charney and Stern, 1962). It thus seems likely that the westward propagating wave packets are baroclinic instability waves steered by the tropical easterly jet. Figure 11d shows the latitude-time section of PV on the virtual potential temperature $\theta_v = 328$ K surface, the lowest isentropic surface that does not intersects the ground. The PV gradient starts to reverse its sign in mid-May, a time that coincides with the initiation of the first organized westward-moving wave packet.

4.2 Wave structure

Figure 12 shows composite structure of the first westward-traveling wave packet, which is obtained

Fig. 11. Premonsoon (May 16–31) conditions averaged within 160–220E: (a) Virtual potential temperature (thick) and zonal wind velocity (thin lines in ms$^{-1}$); (b) Ertel’s isentropic potential vorticity (solid; 10$^{-4}$ K m$^2$kg$^{-1}$s$^{-1}$) and virtual potential temperature (dashed); and (c) meridional circulation as functions of latitude and height. (d) Latitude-time section of potential vorticity on $\theta_v = 328$ K surface, with the area where potential vorticity decreases northward shaded.
by averaging on a coordinate moving westward at 5.5°/day. The wave structure below 500 mb is typical of a baroclinic instability wave in an easterly shear, with a clear up-eastward phase tilt. The cyclonic flows associated with a surface depression advects hot continental (cold marine) air southwestward (northward), giving rise to a net southward heat transport and reducing the meridional temperature gradient near the surface. This westward traveling wave also carries a net northward transport of moisture. East of the surface depression where moist marine air flows inland, a large positive humidity anomaly is found near the ground. Slightly to the east of the surface depression, there is a weak deep convective event that brings about weak precipitation. Intensive southerly winds associated with the baroclinic disturbance funnel moisture inland (Fig. 10c), raising low-level moisture and triggering deep convection on a regional scale.

In contrast to the ground-trapped structure of the first baroclinic wave, the third wave packet that marks the onset of summer monsoon develops a deep structure with large amplitudes (Fig. 13). The lower half of the onset wave still bears the characteristics of a baroclinic instability with a southward heat transport. In the upper troposphere, it has a CISK (conditional instability of the second kind) type structure with convective heating generating a positive temperature anomaly centered on 300 hPa. With strong convective heating, the onset wave can be viewed as a hybrid of baroclinic instability in the lower and CISK in the upper troposphere. In analyzing the stability of an easterly shear, Moorthi and Arakawa (1985) show that latent heating is necessary for such a hybrid mode, with a deep vertical structure, to become unstable (see also Shukla, 1978). How the hybrid wave attains the long period (2 weeks), and long wavelength (10,000 km), needs further theoretical studies.

Precipitation can be very different between baroclinic waves in an easterly and westerly shear. In a westerly shear, precipitation takes place to the east of a cyclone where surface moistening due to the northward advection coincides with the rising motion in the warm sector. In an easterly shear, on the other hand, the surface moistening and rising motion associated with a pure baroclinic instability occur on the eastern and western sides, competing for
precipitation. Thornicroft and Hoskins (1994) suggest that precipitation can occur on either side of a depression in the easterly shear, depending on convective parameterization. In our GCM, deep convection happens to the east of a surface depression, footing on a moist surface layer due to northward advection of marine air. The competition between the rising motion and moisture advection associated with easterly-shear baroclinic instability may make it difficult to organize rainfall and cause the delay in monsoon onset.

4.3 Slow adjustment

The monsoon onset triggered by explosive growth of westward-moving disturbances brings out large changes in flow and temperature fields. Precipitation wets and cools the ground, suppressing dry convection and re-establishing temperature stratification in the lower atmosphere. Associated with this low-level cooling is upper-level latent heating, that reverses the meridional temperature gradient in the upper troposphere, a condition commonly associated with the onset of observed Indian monsoon. With temperature increasing northward throughout the troposphere, the mid-level easterly jet on the coast suddenly relocates itself to the upper troposphere and intensifies (Fig. 14a). Meanwhile, westerlies appear in the low levels. These post-onset adjustments are similar to observations. With wet and cool ground surface near the coast, the baroclinic zone moves northward. The resultant wave disturbances bring rainfall successively further inland, as in Fig. 8.

Baroclinically unstable condition and westward-moving disturbances first appear in mid-May, but what causes the explosive growth of the onset wave in late June? The increasing solar radiation is not the major cause, given the solar radiation does not change much during June as it approaches the solstice. Indeed in the perpetual June 1 run (Fig. 8b), the monsoon still starts in late June as in the seasonal run. Some slow adjustments thus seem involved. To see if the soil moisture adjustment affects monsoon onset, we perform an additional perpetual June 1 run with the soil moisture field fixed at its June 1 values as well (Fig. 8c). Compared to the other two runs with interactive ground hydrology, the first major precipitation event with 140°E–160°W mean rainfall exceeding 2 mm/day takes place on the continent about the same time in the fixed soil moisture run. The reduction in rainfall associated with this onset event is presumably due
to the lack of moisture supply from land surface. Ensemble runs are necessary to determine whether the small difference (~one week) in the onset date among the three experiments is real or a reflection of the monsoon's chaotic aspect.

A slow dynamic adjustment is implied by Plumb and Hou's (1992) transient experiments where it takes a resting atmosphere more than one month to establish a direct circulation in response to a given off-equatorial heating. There is a tendency for the adjustment time to increase with the heating intensity (see their Fig. 6). Solar radiation heats land surface, but does not directly lead to deep convection. The baroclinic instability initiates a deep direct circulation that enhances northward moisture transport. This interaction of deep convection and direct circulation may have led to the eventual explosive growth of the hybrid baroclinic-CISK instability. Consistent with this picture of slow adjustment, the upper troposphere downdraft in northern subtropics starts to weaken in mid-May (Fig. 14b) when the baroclinic instability triggers the first deep latent heating to the atmosphere. The surface westerlies off the coast also first appear at the same time in mid-May (Fig. 14a), indicating a southward surface pressure gradient and hence the first sign of an inland surface flow.

Our experiments do not rule out the possibility that soil moisture adjustment may play a role in delaying the monsoon onset. Further studies are needed to understand slow dynamic and soil moisture adjustments and their roles in preparing conditions for a monsoon's arrival. In this regard, it will be interesting to replace the bucket model with a swap model that has a zero heat content, but allows unlimited water supply for evaporation. Studying such a swap-sea system might help understand the effect of soil moisture adjustment on the onset, and other development characteristics, of a summer monsoon.

Observed pre-monsoon conditions over the Indian subcontinent are similar to those in our model sim-
ulation. Reversed meridional temperature gradient in the lower atmosphere, a mid-level easterly jet, and the local PV minimum due to the deep mixed layer over the heated subcontinent are all present (Fig. 15). In addition to the major minimum extending up-southward from the ground south of 25N, there is a second PV minimum around 35N above 550 mb in observations. The latter northern minimum is presumably caused by elevated heating on the Tibetan Plateau, an effect we will discuss in the next section. Over India, thunder showers prior to the onset of monsoon usually follow the passage of cycloic disturbances (P.N. Vinayachandran and S.K. Behera, 1999, personal communications). This phase relationship between surface pressure, and precipitation, is similar to that of the baroclinic waves in our model simulation.

5. Effects of mountain

Observations and model studies show that elevated heating by the Tibetan Plateau is an important element of the observed Asian monsoons. Here we introduce into the model an isolated topographic feature bounded within 175-243°E and 19°N-47°N. The mountain rises gently to a maximum elevation of 4000 m at (208°E, 33°N), and is symmetric with respect to the summit. The model topography is quite different from the real Tibetan Plateau which rises sharply on its southern flank. Because of this
and other idealizations, the experiment is intended only to get a crude sense of mountain effects. The model is run for two years. We will only show the last year's result as the two summers are grossly similar.

Figure 16 shows the average summer precipitation. A low pressure forms over the mountain in the lower half of the atmosphere, because elevated mountain surface is heated directly by solar radiation. The associated cyclonic circulation advects humid marine air inland on the eastern flank, but dry air southward on the western flank of the mountain. As a result, summer precipitation moves farther northward east of the mountain, while arid conditions persist to the west—a rainfall distribution consistent with observations. West of the topography, the monsoon rainbelt is pushed slightly southward and now stays off the coast over the ocean, in support of the monsoon-desert mechanism discussed in Rodwell and Hoskins (1996). Compared to Fig. 4, the mountain-induced precipitation changes are much larger to the east than to the west of the mountain. This east–west asymmetry of the mountain effect further underscores the importance of moisture transport in determining summer climate. Baroclinically unstable pre-monsoon conditions similar to Fig. 13 exist in simulation with the mountain. The westward propagating waves amplify much earlier, probably because the mountain raises the baroclinically unstable stratification to higher altitudes, and supplies finite amplitude initial disturbances. As a result, the NH monsoon rain belt appears one month earlier in late May/early June (Fig. 17). West of the mountain, the rain belt penetrates less far inland than in the no-mountain case, leaving arid conditions to the western continent. Both the downwelling and the low-level southward flows associated with the mountain-forced Rossby waves (Rodwell and Hoskins, 1996) might contribute to this drying of the western continent.

6. Discussion

GCM experiments are conducted with an idealized land-sea distribution to investigate the processes and mechanisms controlling the formation of subtropical deserts and monsoon. We focus on summer precipitation because it determines the annual
mean precipitation over the subtropics where the subsidence associated with the Hadley circulation suppresses rainfall in other seasons. The bucket depth in the simple hydrolology model is set constant and the land surface albedo and SST are zonally uniform. So the land-sea distribution is the only zonal asymmetry externally imposed and there is no a priori preference for one longitudinal sector to be wet or dry over the other. The distributions of precipitation and hence climate type are thus determined solely by atmospheric circulation, and its interaction with land hydrological cycle.

A gigantic high pressure system over the ocean and a heat low over the subtropical continent are the dominant features of the summer surface circulation. The associated southerly winds help transport moisture inland, favoring a wet monsoonal climate in the eastern continent. In the western continent, the northerly winds advect dry continental air and push rainbelt southward. In a seasonal experiment with flat continental surface, this pair of continental low and oceanic high maintains a zonal asymmetry in precipitation distribution over the subtropical continent, with a wet climate in the east and deserts in the west.

The model climate in a perpetual summer integration looks drastically different from the one presently observed, featuring monsoon rainbelt that penetrates so deep inland as to cover the present arid Sahara and Arabian deserts. We show that the slow interaction between rainfall and soil moisture is responsible for the differences between the seasonal and perpetual runs. The soil moisture distribution dictates where rain falls over land in summer. Dry ground heats up in summer and precludes deep moist convection because dry convection redistributes moisture over a deep mixed layer and depletes moisture near the ground. By contrast, wet surface favors deep convection, trapping moisture within a shallow mixed layer and providing local supply of moisture via evaporation. In perpetual summer runs, the monsoon rainbelt migrates slowly northward to 30°N in association with the surface wet zone. Because its northward migration is much slower than the sun’s, the rainbelt cannot penetrate very far inland before the solar radiation starts to decrease, pulling it back toward the south. This premature withdrawal of monsoon rainbelt leaves the western subtropical continent without rain in the seasonal model run. Thus the summer atmosphere is not in equilibrium with changing insolation, a fact overlooked in literature, but key to the formation of subtropical deserts. Because of the unsteady nature of the summer climate, and because the speed of the monsoon rainbelt migration depends on the strength of the hydrological cycle, the semi-arid Sahel region separating the wet equatorial Africa from deserts to the north is particularly prone to variability.

On geological time scales, solistic insolation varies. When the seasonal solar forcing becomes stronger as in the mid-Holocene, the precipitation associated with the African monsoon is likely to increase in intensity and advance to higher latitudes (Joussaume et al., 1999), because increased rainfall will accelerate the northward migration of the monsoon rainbelt. This mechanism, along with others like the interaction with the South Asian monsoon (Rodwell and Hoskins, 1996) and soil/plant type (Kutzbach et al., 1996), may be responsible for the large variability in both the intensity and northern reach of the African monsoon.

The observed Asian summer monsoons are characterized by their delayed abrupt onset. Our idealized GCM simulation demonstrates that these onset characteristics are shared by a summer monsoon in its simplest form due to land-sea distribution alone. Our analysis of the simulation suggests the following sequence of development that eventually leads to the onset of monsoon. After the spring equinox, increasing solar radiation heats up the land surface very quickly, while the ocean surface to the south remains cool because of greater heat content of the ocean mixed layer. On a non-rotating earth, the cool marine air would begin to intrude inland. The moisture it carries would condense into monsoon rain to sustain the sea breeze winds. In a rotating atmosphere, the land-sea temperature difference does not necessarily lead to a direct overturning cell, but instead can be balanced by adjusting its vertical shear into a thermal wind equilibrium. The simulated monsoon onset is associated with the explosive growth of a westward traveling disturbance that displays a deep vertical structure indicative of baroclinic instability in the lower and CISK in the upper troposphere.

Two conflicting views of monsoon exist. On one hand, the 2D sea breeze model is an often used metaphor for summer monsoon, especially when it comes to explain its rainfall, which does not mention Earth’s rotation and obviously fails to explain monsoon’s delayed onset. Geostrophic/thermal wind balance, on the other hand, is often invoked to explain monsoonal winds with a given temperature distribution, but its implications for monsoon’s temporal development have not been made clear. The sequence of development described in the last paragraph offers a new conceptual model to describe the monsoon in a self-consistent way. We call it “Geostrophic Monsoon” because geostrophic balance is of central importance, directly responsible for the delay of its onset despite the large land-sea temperature difference. Furthermore, the first deep convective events that initiate the deep direct circulation are triggered by quasi-geostrophic disturbances due to the instability of the thermal wind balance. Much of the inland moisture transport is carried out by boundary layer flows below 850 mb,
which contain a large ageostrophic component due to friction. This ageostrophic aspect of monsoon does not necessarily contradict the geostrophic monsoon concept, as one can argue that the boundary layer flows themselves are largely controlled by the pressure field in geostrophic balance aloft. For example, the moisture transport events in Fig. 10c are clearly associated with quasi-geostrophic baroclinic instability waves, although much of the moisture itself is carried inland by ageostrophic boundary layer flows.

Further model experiments under controlled conditions can provide direct tests of some key elements of the proposed geostrophic monsoon concept. For example, the geostrophic control of monsoon onset can be tested by changing the rotation rate in the GCM and seeing how the onset date might vary. By running an axisymmetric version of the GCM that eliminates the triggering disturbances, we might see a later onset or the disappearance of monsoon altogether. In our GCM simulation, shallow baroclinic wave packets with light precipitation proceed the final onset by as much as one month, suggesting that some slow processes other than solar radiation are involved. Further work is needed to identify and understand the slow adjustment that leads to the abrupt onset of monsoon.

Despite the heavy idealization of our GCM setup, simulated monsoon displays several features reminiscent of observations. There is a PV minimum over the pre-monsoon Indian Subcontinent, indicating the possibility of baroclinic instability. Biweekly oscillations are observed, and at least some of them originate from the western Pacific and propagates westward. Maps of climatological onset dates indicates a northwestward propagation from the Southeast Asia through Bay of Bengal to India (Murakami and Matsumoto, 1994; Wang, 1994). The individual onset seems more complicated, with the baroclinic instability as but one of many possible mechanisms. Such complexities of the real monsoon may be attributed to those in coastal like, SST, vegetation and topography. As a first step to close these gaps, we investigate the effects of Tibetan Plateau. Rainfall moves farther inland to the east, but retreats southward to the west of the mountain. These precipitation changes are consistent with a cyclonic circulation induced by the elevated mountain surface heating. Further insights can be gained by adding to the model more realistic features of the land-sea distribution. The results here can serve as a reference against which more realistic simulations are compared to elucidate the effects of added new features.

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理想化海陸分布を持つ大気大循環モデルに現れる
夏期モンスーンの急な開始およびその季節進行

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世界最大の砂漠と最大のモンスーン域は共にほぼ同じ緯度のアフリカ・ユーラシア大陸の亜熱帯に存在する。このような降水の東西分布の形成機構を調べるために、大気大循環モデルを用いて理想化した海陸分布の下で実験を行った。特にここではアジア大陸の南岸を北緯17度線に沿って東西真っ直ぐに設定した。

山岳がなく、また海面水温と陆面パラメータが東西一様であるにも関わらず、夏期の降水は亜熱帯大陸の東部で多く、西部で少ない。このような降水分布の東西非一様性は夏期の海洋上に現れる高気圧の水蒸気輸送によると考えられる。更に、降水と土壌水分のゆっくりとした相互作用はモンスーン降水帯の北進を遅らせ、土壌水分が十分に増加する前に、太陽放射の着熱によって降水帯は南下し、大陸内部まで進出できない。太陽放射を夏の値に固定した実験では、モンスーン降水帯が徐々に北進し、北アフリカ全域をカバーするようになった。

モデルのモンスーン降水は6月後半に突然大陸南岸に現れる。このようなモンスーンの急な開始は西進する波動の発達に乗って起きる。春分後、暑い大陸と冷たい海洋間には北向きの温度差が大気下層で形成され、時間と共に強化されている。この下層の南北温度差は偏東風シアーとほぼ温度差バランスをし、背の高い南北モンスーン循環は形成されない。しかし、このような温度差バランスを最終的には傾圧的に不安定になり、湿潤傾圧不安定の爆発的成長によってモンスーンが始まると。

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