The relative impacts of various land–sea distributions (LSDs) and mountains on Asian monsoon extent and intensity are assessed using a series of AGCM simulations. The air–sea coupling effects are not considered in this study. All simulations were integrated with zonal mean SST, globally uniform vegetation, soil color, and, except several simulations, soil texture. The results show that the LSD plays a more fundamental role than orography in determining the extent of Asian and African monsoons. The tropical zonal LSD and Asian mountains both play a crucial role for establishing summer monsoon convection over the South Asian region. The monsoon circulation index (MCI1) defined by the difference of zonal wind between 850 and 200 hPa is used to measure the intensity of the South Asian summer monsoon. The large-scale meridional land–sea thermal contrast between the Eurasian continent and the Indian Ocean only induces a 1.8 m s\(^{-1}\) increase of MCI1. The presence of the Indian subcontinent and Indochina peninsula (Asian mountains), however, induces a 6.6 (7.4) m s\(^{-1}\) increase of MCI1 associated with the release of latent heat of condensation. Clearly, the tropical subcontinental-scale zonal LSD and the Asian mountains almost equally contribute to the increase of MCI1 and play a more important role than the large-scale meridional LSD between the Eurasian continent and the Indian Ocean. Possible mechanisms of how the tropical subcontinental-scale zonal LSD and Asian mountains impact the Asian summer monsoon circulation and precipitation are also discussed.

1. Introduction

Land–sea distribution (LSD), orography, and land cover strongly impact the monsoon circulation and precipitation by altering the exchange of energy, momentum, and water between the underlying surface and the atmosphere. LSD and orography play different roles in the Asian monsoon system. The effects of an idealized LSD and the Tibetan Plateau on the Asian monsoon circulation have been discussed in Dirmeyer (1998), Chou (2003), and Liang et al. (2005, 2006). They found that a model can capture the major features of the Asian summer monsoon when an idealized LSD (e.g., Eurasia, Africa, the Indian subcontinent, and Indochina peninsula) were included in the experiment. When the effect of the Tibetan Plateau was considered, the model produced an intensified South Asian summer monsoon and East Asian summer monsoon characterized by an intensified meridional temperature gradient over South Asia and a northward extension of the Asian summer monsoon rain belt. Effects of Tibetan Plateau uplift were discussed in Kutzbach et al. (1989), Prell and Kutzbach (1992), and An et al. (2001). Ramstein et al. (1997) simulated the Eurasian climates of today, 10 million, and 30 million years ago by using an AGCM that incorporates realistic continental geography and epicontinental sea distribution. Their results suggested that the retreat of the Paratethys (an epicontinental sea) played a role just as important as the uplift of the Tibetan Plateau in driving Asian monsoon changes.

In addition, the land surface condition, such as soil, vegetation, and snow, also affects the magnitude of monsoons (Tzeng and Lee 2001; Qian et al. 2003; Wu and Qian 2003; Xue et al. 2004). Yasunari et al. (2006) evaluated the relative roles of the Tibetan Plateau and continental-scale land surface processes (including soil and vegetation) in
the Asian monsoon by using an AGCM. Their results showed that the land surface and Tibetan Plateau effects contribute nearly equally to increased precipitation in the South Asian and East Asian monsoon regions.

Although many studies have been focused on the role of LSD, orography, or the land surface effect in the Asian monsoon system (e.g., Halley 1686; Hahn and Manabe 1975; Ye and Gao 1979; Dirmeyer 1998; Ren and Qian 2002; Liu and Yin 2002; Kitoh 2004; Liang et al. 2005; etc.), there still exist arguments as to the origin of monsoons. Which role is more fundamental in the formation of the Asian monsoon—the LSD or orography? An online open discussion had been initiated by M. Yanai on the origin of monsoons in 2001. However, no general consensus was achieved among all scientists on this argument. Some studies suggested that the LSD is fundamental in formation of the Asian monsoon (Kuo and Qian 1982; Webster 1987); some emphasized the importance of the Tibetan Plateau (Prell and Kutzbach 1992; Wu and Zhang 1998; Liu and Yin, 2002; Wu et al. 2007). Chao and Chen (2001) suggested that LSD and orography are not necessary for the Asian and Australian monsoon because most features of the Asian and Australian summer monsoon circulation can be reproduced even without Asia, maritime continents, and Australia in the numerical experiment. We believe that a systemic evaluation of the relative impacts of various LSDs and mountains on the Asian monsoon is necessary. The response of the atmospheric circulation to various LSDs and mountains could be very different owing to differences in location and size of the LSD and mountain.

We do not intend to determine here whether LSD or orography has a more fundamental impact on the Asian monsoon system; rather, we will provide an objective assessment on the relative contributions of various LSDs and mountains to the Asian monsoon from different aspects. It should also be emphasized that this study does not aim at performing a paleoclimate simulation and understanding formation of the paleoclimate at a given geological time but our focus is to identify the response of the atmosphere to various LSDs and mountains and to explore their mechanisms. However, it is still helpful for understanding the paleoclimate in different periods of the earth’s history. A coupled ocean–atmosphere GCM (OAGCM) should provide a more realistic simulation of the effects of LSD and orography, while significant modifications of the LSD and orography in a numerical model will make it harder to control stability of the numerical integration. In the current study, we do not consider air–sea coupling effects.

The organization of this paper is as follows: A brief description of the model and experimental design is presented in section 2. Section 3 presents the relative contributions of various LSDs and mountains to the monsoon extent, intensity, and atmospheric heat sources and moisture sinks. Section 4 elucidates the possible mechanisms of the Asian monsoon response to changes of LSD and orography. A summary and discussion are given in section 5.

2. Model and experimental design

The model used is the Community Climate Model, version 3 (CCM3), developed by the National Center for Atmospheric Research, which is a global spectral climate model with standard resolution T42 (approximately a 2.8° × 2.8° transform grid) and 18 layers in the vertical. The process of deep convection is treated with a parameterization scheme developed by Zhang and McFarlane (1995). The CCM3 incorporates a one-dimensional land surface model of energy, momentum, water, and CO2 exchange between the atmosphere and land (Bonan 1998). The land surface model includes 12 plant types that form 28 different vegetation surfaces. Soil effects are included by allowing thermal and hydraulic properties to vary depending on percentages of sand and clay. Soils are divided into eight kinds of colors to define saturated and dry soil albedos (Bonan 1996). A complete description of the physical and numerical methods used in CCM3 was provided by Kiehl et al. (1996, 1998a). CCM3 gives an adequate reproduction of the mean climate state, notwithstanding some biases and errors in comparison with observations. Detailed analyses and comparisons of the climate simulation of CCM3 to various observations were provided by Hurrell et al. (1998) for the dynamical simulation, Hack et al. (1998) for the hydrologic and thermodynamic simulation, and Kiehl et al. (1998b) for the energy budget simulation. Over the past several years, CCM3 has been used by many scientists for climate research into such areas as CO2 warming, climate change, paleoclimate, climate prediction, and predictability.

To accentuate the effect of land–sea contrast, inhomogeneities of the land surface were removed by using a globally uniform vegetation type, soil color, and soil texture. We used “warm grassland,” covering the largest area of land surface, as the prescribed land surface type, and “4” as the parameter of soil color that also covers the largest area of land surface among various soil colors. The uniform soil texture was prescribed with the global average fraction of sand (50.6%), silt (25.7%), and clay (23.7%). To remove information of the present-day LSD in the initial conditions and the external forcing, the zonal mean initial fields and prescribed zonal mean SST were used. Another consideration in using the zonal mean
SST is to isolate the effect of zonal land–sea thermal contrast from the zonal difference of SST. As will be seen from section 4a, most low pressure systems and heavy rainfall are created over tropical continents in summer and then propagate westward, indicating the importance of tropical continents in the establishment of heavy monsoon rainfall. These features are not very clear in observations or in the simulation with realistic SST because the zonal land–sea thermal contrast is mixed with the zonal difference of SST.

It is well known that the pattern of SST strongly impacts monsoon circulation. The Asian monsoon circulation can be simulated by a numerical model with realistic SST distribution even without land–sea thermal contrast (Chao and Chen 2001). What would happen if we retain the land–sea thermal contrast but remove the zonal inhomogeneity of SST? The zonal mean SST simulations are also helpful to answer this question. Here the zonal mean SSTs were calculated by using the observed climatological monthly SSTs (Shea et al. 1992). Each experiment was integrated 13 years with the zonal mean SSTs. The pentad mean and monthly mean outputs from the last 10 years of simulations were then saved for analysis.

A series of numerical experiments were designed to explore the effects of various LSDs, mountains, SSTs, and land surface conditions on the Asian monsoon. Table 1 provides a summary of the experiments discussed in this paper. These experiments can be grouped into the following four categories with each category consisting of various runs.

1) **Land–sea distribution experiments**: Various flat continents were introduced one after another from the aquaplanet (AP) run only with zonal mean SST distribution to the NM run with realistic LSD (Table 1). The LO32, LO21, and LO10 runs serve to explore the role of large-scale meridional LSD at different latitudes in the Asian monsoons. Comparison among the POAO, AfrSAm, India, and SAsia runs helps understanding the role of the zonal LSD, as the presence of a tropical continent or peninsula, in the Asian monsoon.

2) **Mountain experiments**: The Asian and African mountains were introduced one after another based on the NM run to identify the effect of these regional mountains on the Asian monsoon. The runs in categories 1 and 2 were integrated with zonal mean SST.

3) **SST experiment**: In this category, the M2 run is the same as the M run except that the zonal-mean SST was replaced by realistic climatology to identify the effect of a realistic SST distribution on the Asian monsoon. The runs in categories 1–3 were integrated with a globally uniform vegetation type, soil color, and soil texture as described previously.

4) **Control experiment**: The control run (Ctrl) is the same as the M2 run except that the globally uniform vegetation type, soil color, and soil texture were replaced by realistic properties. The Ctrl run is mainly used to diagnose the performance of the CCM3 in monsoon simulation by comparing with observations.

<table>
<thead>
<tr>
<th>Category</th>
<th>Run</th>
<th>Underlying surface of the model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Land–sea distribution experiments</td>
<td>AP</td>
<td>The whole planet is covered by ocean with zonal mean SST except that south of 78°S is covered by sea ice</td>
</tr>
<tr>
<td></td>
<td>LOn</td>
<td>As in AP but the area north of n°N is covered by land with globally uniform vegetation type, soil color, and texture; n = 32, 21, 10 (see Fig. 1a for the LO21 run)</td>
</tr>
<tr>
<td></td>
<td>POAO</td>
<td>Pacific and Atlantic Oceans are included based on LO21. (Fig. 1b)</td>
</tr>
<tr>
<td></td>
<td>AfrSAm</td>
<td>African and South American continents are included based on POAO. (Fig. 1c)</td>
</tr>
<tr>
<td></td>
<td>India</td>
<td>Indian subcontinent is included based on AfrSAm (Fig. 1d)</td>
</tr>
<tr>
<td></td>
<td>SAsia</td>
<td>Indochina peninsula is included based on the India run (Fig. 1e)</td>
</tr>
<tr>
<td></td>
<td>NM</td>
<td>Present-day land–sea distribution, but with no orography (Fig. 1f)</td>
</tr>
<tr>
<td>Mountain experiments</td>
<td>MAsia</td>
<td>Asian large-scale mountains (Tibetan, Iranian, and Mongolian Plateaus) are included based on NM (Fig. 1g)</td>
</tr>
<tr>
<td></td>
<td>MAsiaAfr</td>
<td>African mountains are included based on MAsia (Fig. 1h)</td>
</tr>
<tr>
<td></td>
<td>M</td>
<td>Present-day land–sea distribution and orography (Fig. 1i)</td>
</tr>
<tr>
<td>SST experiment</td>
<td>M2</td>
<td>As in the M run but with realistic climatological SST distribution, varying monthly</td>
</tr>
<tr>
<td>Control experiment</td>
<td>Ctrl</td>
<td>As in M2 but with realistic vegetation type, soil color, and soil texture distribution</td>
</tr>
</tbody>
</table>

3. Relative contributions of various LSDs and mountains to the Asian monsoon

a. Extent of the Asian monsoon

Previous studies mainly focused on the influences of LSD or orography on the monsoon intensity by diagnosing
the monsoon circulation, rainfall, and so on (e.g., Li and Yanai 1996; Dirmeyer 1998; Chou 2003). Little work has been devoted to evaluating the influence of LSD and orography on the extent of the Asian monsoon. As the word “monsoon” indicates, the Asian monsoon is part of a seasonally reversing wind system (Ramage 1971; Rao 1976) characterized by wet summers and dry winters. To measure the extent of monsoons, a monsoon precipitation index (MPI)

$$\text{MPI} = \frac{R_S - R_W}{R_M},$$

defined by Wang and Ding (2008), was calculated by using the climatological precipitation of each experiment; $R_S$ and $R_W$ represent the mean monthly precipitation in the summer monsoon season (May–September) and winter monsoon season (November–March) in the NH and $R_M$ is the mean monthly precipitation throughout the year. The annual range of precipitation is defined by the difference between the local summer and winter precipitation ($R_S - R_W$) in the NH or SH. The monsoon precipitation domain (MPD) is determined by MPI > 0.5 and annual range of precipitation greater than 60 mm month$^{-1}$.

Figure 2 shows the MPDs and the regions with the seasonally reversing wind system, defined by the angle between summer [June–August (JJA)] and winter [December–February (DJF)] prevailing wind larger than 90° for observation and various runs. The observed MPD is calculated by using the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) and the NCEP reanalysis wind. The other MPDs are based on the model outputs. Comparison of Figs. 2a and 2b shows that the simulated MPDs and regions with the seasonally reversing wind system are in good agreement with observations. Both clearly reveal monsoon rainfall in South Asia, West Africa, and North Australia.
However, the model cannot reproduce well the East Asian subtropical monsoon domain owing to poor simulation of the East Asian monsoon rainfall.

In the AP run and LO32 run, almost no MPD can be found in the NH (not shown); however, the MPD starts to appear over the southern part of the continent in the LO21 run and extends southward to cover the low latitudes from 5° to 30°N entirely in the LO10 run (Figs. 2c,d). In addition, a MPD can be found over austral oceans around 15°S due to the seasonal march of heavy rainfall associated with the meridional migration of solar radiation and zonal mean SST (Fig. 2c). The oceanic monsoon region can also be found in other LSD experiments, such as the AP, POAO, AfrSAm, and SAsia runs, and shows a smaller MPI (MPI < 0.5) than that in the continental monsoon region (e.g., Figs. 2d–g). The intense monsoon characterized by the larger MPI (MPI > 1) starts to occur when the tropical continent or peninsulas are introduced. For example, a MPD covers northern Africa associated with large MPI and the seasonally reversed wind when the African and South American continents are introduced based on the POAO run (Figs. 2e,f). The presence of the Indian subcontinent and Indochina peninsula expands the MPD southward from 20°N to the near-equatorial region over South Asia (Figs. 2f,g).

Comparison of the MAsia run with the NM run shows that the presence of Asian mountains helps to extend the seasonally reversing wind system northeastward over East Asia, consistent with the findings of Liu and Yin.

Fig. 2. The monsoon precipitation index (contours) within the monsoon precipitation domains (thick black curves), and the regions with a seasonally reversing wind system at 925 hPa (shading) for various experiments. The monsoon precipitation domain is defined by MPI > 0.5 and the annual range of precipitation is greater than 60 mm month⁻¹. Gray dots indicate orography.
(2002). However, the MPD is absent south of 35°N, as in the Ctrl run, owing to poor simulation of the East Asian rainfall (Figs. 2b,i). The other mountains have little influence on the extent of Asian–Australian–African monsoons (Figs. 2j,k). When the zonal mean SST is replaced by a realistic climatological mean, the seasonally reversing wind system starts to occur in Southeast Asia, Indonesia–Australia, and the southern equatorial Indian Ocean. However, the response of the MPD is relatively insensitive to SST change except over the South China Sea and the Australian monsoon region (Figs. 2k,l).

Here, we define the MPDs with a seasonally reversing wind system at 925 hPa as the “typical monsoon regions.” Table 2 presents the area percentage of typical monsoon regions in various numerical experiments relative to that in the Ctrl run. In the tropical Asian–Australian–African monsoon region (25°S–25°N, 30°W–160°E), the area of typical monsoon regions expands from 12% in the LO21 run to 197% in the LO10 run due to the southward extension of the continent from 21° to 10°N, indicating the crucial role of a tropical continent in determining monsoon extent. The area of a typical monsoon region expands from 8% in the POAO run to 68% in the AfrSAm run and further to 98% in the SAsia run. Clearly, the tropical continents—even the Indian subcontinent and Indochina peninsula—play a crucial role in determining the extent of a typical monsoon region.

In contrast, the monsoon area shows a 19% increase after the global orography is introduced, which is smaller than the increase (30%) induced by the presence of the Indian subcontinent and Indochina peninsula. In addition, the monsoon area expands from 96% in the M run to 109% in the M2 run following the replacement of the zonal mean SST by a realistic one. In the tropical Asian–Australian monsoon region (25°S–25°N, 50°–160°E), we can draw similar conclusions as to the contribution of LSDs and mountains to the monsoon extent, except that influence of the African and South American continents on the Asian monsoon extent is negligible. The reason is that the Asian monsoon region is located to the east of the African and South American continents; however, the increasing precipitation and westerly wind induced by the presence of tropical continents occur to the west of the continent. This will be discussed further in section 4. In addition, the 37% increase of monsoon area is found by comparing the M2 run with the M run, which mainly results from the appearance of the South China Sea monsoon and Australian monsoon in the M2 run (Figs. 2k,l).

b. Asian summer monsoon circulation and precipitation

1) LONGITUDE–PRESSURE SECTION OF ZONAL WIND

The large-scale South Asian summer monsoon circulation is characterized by a deep baroclinic structure with low-tropospheric westerlies and upper-tropospheric easterlies. To explore the contributions of various LSDs and mountains to the South Asian zonal vertical circulation, Fig. 3 illustrates a longitude–pressure section of the summer (June–September) zonal wind component averaged over the latitude belt from 10° to 20°N. Here we only show the longitudinal range between 20°W and 160°E. The dark shadings denote that the westerlies (easterlies) in the current run are significantly stronger (weaker) than that in the reference run. By contrast, the light shadings denote that the easterlies (westerlies) in the current run are significantly stronger (weaker) than that in the reference run. Square brackets indicate the reference run.

The Ctrl run can reasonably capture primary features of the zonal wind, such as the low-tropospheric westerlies and upper-tropospheric easterlies over South Asia (Figs. 3a,b). In the AP run, the zonal wind appears to be very weak over the latitude belt from 10° to 20°N (Fig. 3c). With the presence and southward extension of the boreal continent, the low-tropospheric westerlies start to occur and intensify accompanied with the intensification of midupper-tropospheric easterlies (Figs. 3d,e). The comparison of the AfrSAm run with the POAO run indicates that the presence of the African and South American continents induces a westerly wind below 500 hPa and an intensified easterly wind above 300 hPa over the longitude belt of Africa as well as the adjacent Atlantic Ocean (Figs. 3f,g). A similar result can be found over South Asia when the Indian subcontinent and Indochina peninsula are introduced, although the size of the subcontinent

<table>
<thead>
<tr>
<th>AP</th>
<th>LO32</th>
<th>LO21</th>
<th>LO10</th>
<th>POAO</th>
<th>AfrSAm</th>
<th>SAsia</th>
<th>NM</th>
<th>MAsia</th>
<th>MAsiaAfr</th>
<th>M</th>
<th>M2</th>
</tr>
</thead>
<tbody>
<tr>
<td>AAA</td>
<td>0%</td>
<td>0%</td>
<td>12%</td>
<td>197%</td>
<td>8%</td>
<td>68%</td>
<td>98%</td>
<td>77%</td>
<td>78%</td>
<td>91%</td>
<td>96%</td>
</tr>
<tr>
<td>AA</td>
<td>0%</td>
<td>0%</td>
<td>12%</td>
<td>174%</td>
<td>11%</td>
<td>12%</td>
<td>48%</td>
<td>31%</td>
<td>38%</td>
<td>58%</td>
<td>63%</td>
</tr>
</tbody>
</table>
and peninsula is smaller than the African continent (Fig. 3h). Clearly, the tropical continent helps to build or strengthen the low-tropospheric westerlies and upper-tropospheric easterlies over the continent and to its west. In both LO10 and SAsia runs, the southernmost continental extent is 10°N for the South Asian region. The tropical continental area is larger in the LO10 run than that in the SAsia run. However, the westerlies extend upward to near 600 hPa in the SAsia run but are below 800 hPa in the LO10 run, which suggests that the zonally asymmetric heating induced by the tropical zonal LSD is crucial for establishing deep convection, which in turn intensifies the South Asian monsoon westerlies.

Presence of the Asian mountains also significantly strengthens the South Asian westerlies and extends them
upward from 850 hPa in the NM run to 500 hPa in the MAsia run; meanwhile, the upper-tropospheric easterlies appear to be significantly strengthened in association with the intensification of the South Asian low in the low troposphere and the South Asian high in the upper troposphere due to the sensible heat and release of latent heat of condensation over the Tibetan Plateau (Figs. 3i,j). This will be discussed in detail in section 4b. The presence of African mountains helps to further strengthen the low-tropospheric westerlies and upper-tropospheric easterlies over South Asia (Figs. 3j,k). The zonal difference of SST is introduced when the zonal mean SST is replaced by a realistic one. As a result, both low-tropospheric westerlies and upper-tropospheric easterlies are significantly strengthened over a broader area (Fig. 3l). The zonal heating contrasts induced by the SST and LSD likely play a similar role in the formation of tropical deep convection.

2) Vertical Shear of Zonal Wind and Monsoon Rainfall

The vertical shear of zonal wind and monsoon rainfall have been widely used to measure the strength of the South Asian summer monsoon. The monsoon circulation index (MCI1), defined by Wang and Fan (1999) using the difference of zonal wind ($U_{850} - U_{200}$) averaged over the Arabian Sea–India region ($5^\circ$–$20^\circ$N, $40^\circ$–$80^\circ$E), is a good indicator of Asian summer monsoon intensity since the vertical shear of zonal wind is closely related to the Asian summer monsoon rainfall as well as the land–sea heating contrast. Table 3 shows the MCI1 and area-mean South Asian summer monsoon rainfall (SAMR) for various runs to explore the relative contributions of various LSDs and mountains to the intensity of the Asian summer monsoon. The SAMR is averaged over the India–Bay of Bengal region (10°–25°N, 70°–100°E) because the intense summer monsoon convection occurs in this region (Wang and Fan 1999).

In the AP and LO32 runs, MCI1 and the seasonality of precipitation appear to be weak (MPI < 0.5) although plentiful precipitation is simulated. The MCI1 increases significantly with the southward extension of the continent from 32° to 21°N and from 21° to 10°N. The SAMR shows a significant increase from 6.7 mm day$^{-1}$ in the LO21 run to 9.6 mm day$^{-1}$ in the LO10 run. These suggest that the tropical continent plays a very important role for establishing summer monsoon circulation and rainfall. Similar results can be found in Dirmeyer (1998). Comparison of the AfrSAm run with the POAO run and with the SAAsia run shows that the presence of the African and South American continents weakens the MCI1 and SAMR, although not significant at the 99% confidence level. However, the presence of the Indian subcontinent and Indochina peninsula leads to a 6.6 m s$^{-1}$ increase in MCI1 and 2.4 mm day$^{-1}$ increase in SAMR. The reason will be elucidated in section 4 based on theoretical studies (e.g., Matsuno 1966; Webster 1972; Gill 1980).

When the Asian mountains are introduced, the MCI1 increases from 4.5 m s$^{-1}$ in the NM run to 11.9 m s$^{-1}$ in the MAsia run due to the heating over the Tibetan Plateau; meanwhile, the SAMR increases from 8.6 mm day$^{-1}$ in the NM run to 10.2 mm day$^{-1}$ in the MAsia run. The presence of African mountains also helps to increase the MCI1 and SAMR in association with the intensification of the Somalian low-level jet. When the zonal mean SST is replaced by a realistic one, both the MCI1 and SAMR are significantly strengthened. The increase of MCI1 induced by the large-scale meridional LSD between the Eurasian continent and the Indian Ocean (POAO–AP), the tropical subcontinental-scale zonal LSD (SAAsia–AfrSAm), the Asian mountains (MAsia–NM), and the realistic SST distribution (M2–M) are 1.8, 6.6, 7.4, and 9.1 m s$^{-1}$, respectively. Clearly, the tropical subcontinental-scale zonal LSD, the Asian mountains, and the realistic SST play a more important role than the large-scale meridional LSD between the Eurasian continent and the Indian Ocean in determining vertical shear of the zonal wind over South Asia.

c. Atmospheric heat sources and moisture sinks

Atmospheric heat sources and moisture sinks play a crucial role in driving the monsoon circulation (Bhide et al. 1997; Ose 1998; Wang and Qian 2000). The vertically integrated apparent heating ($Q_1$) and apparent moisture sink ($Q_2$) can be written as (Yanai et al. 1973; Yanai and Tomita 1998).
\[ \langle Q_1 \rangle = R + LP + SH, \]  
\[ \langle Q_2 \rangle = L(P - E), \]

where \( R \) is the vertically integrated radiative heating rate; \( P, SH, \) and \( E \) are the precipitation rate, the sensible heat flux, and the evaporation rate, respectively; and \( L \) is the latent heat of condensation. Figure 4 shows the vertically integrated apparent heating \( \langle Q_1 \rangle \) and apparent moisture sink \( \langle Q_2 \rangle \) in watts per square meter averaged over the Tibetan Plateau, South Asia, and South China Sea region for various runs.

In the AP and LO32 runs, \( \langle Q_1 \rangle \) and \( \langle Q_2 \rangle \) appear to be negative over the Tibetan Plateau region owing to the intense evaporation in the subtropical region, indicating an atmospheric heat sink and moisture resource (Fig. 4a). In the AP, LO32, and LO21 runs, the atmospheric heat sink weakens with the southward extension of the subcontinent and becomes an atmospheric heat source in the LO21 run. The atmospheric heat source does not show significant change with further southward extension of the continent from 21°N to 10°N. The subtropical atmospheric heat source mainly results from the sensible heat because the sensible heat is about two times greater than the latent heat of condensation. Also, \( \langle Q_2 \rangle \) becomes weak but still appears to be negative in the LO21 and LO10 runs over the Tibetan Plateau region; \( \langle Q_1 \rangle \) and \( \langle Q_2 \rangle \) are relatively insensitive to the presence of the Pacific and Atlantic Oceans, the African and South American continents, and the Indian subcontinent and Indochina peninsula. With the presence of the Asian mountains, \( \langle Q_1 \rangle \) (\( \langle Q_2 \rangle \)) dramatically increases from 43 (−4) W m\(^{-2}\) in the NM run to 149 (94) W m\(^{-2}\) in the MAasia run. Clearly, Asian mountains play a dominant role in heating the air aloft through sensible heat exchange and the release of latent heat of condensation. The influences of other mountains, SST distribution, and land surface properties on \( \langle Q_1 \rangle \) and \( \langle Q_2 \rangle \) are not significant at the 99% confidence level over the Tibetan Plateau region (Fig. 4a).

In the South Asian region of the AP, LO32, and LO21 runs, \( \langle Q_1 \rangle \) and \( \langle Q_2 \rangle \) appear to be relatively weak (Fig. 4b). In the LO10 run, \( \langle Q_1 \rangle \) and \( \langle Q_2 \rangle \) increase respectively by 58 and 25 W m\(^{-2}\) relative to those in the LO21 run. Comparatively, \( \langle Q_1 \rangle \) and \( \langle Q_2 \rangle \) increase by 72 and 46 W m\(^{-2}\) when the Indian subcontinent and Indochina peninsula are introduced (SAasia–AfrSAM). In both LO10 and SAasia runs, the southernmost continental extent is 10°N in the South Asian region. Clearly, the presence of the tropical zonal LSD results in a stronger atmospheric heat source and moisture sink than a land surface without zonal LSD. Comparison of the MAasia run with the NM run suggests that the presence of Asian mountains also helps to intensify atmospheric heat sources and moisture sinks in the South Asian region. The presence of a realistic SST distribution further enhances the atmospheric heat sources.

The vertical profiles of area-mean \( Q_1 \) and \( Q_2 \) are calculated in the South Asian region using pentad-mean wind, air temperature, and moisture fields for various simulations (Fig. 5). In the SAasia run (Fig. 5a), the area-mean \( Q_1 \) has a maximum around 450 hPa; the area-mean \( Q_2 \), however, has its principal maximum around 800 hPa and a secondary peak around 350 hPa. The vertical
profiles of $Q_1$ and $Q_2$ in the MAsia run and the M2 run are similar to those in the SAsia run but with larger heating rates (Figs. 5b,c). These profiles of $Q_1$ and $Q_2$ are similar to those calculated by Yanai and Tomita (1998), indicating that the release of latent heat is associated with cumulus convection (e.g., Yanai et al. 1973).

In the South China Sea region, the changes of $h_{Q_1}$ and $h_{Q_2}$ from the AP run to the LO10 run appear to be similar to those in the South Asian region (Figs. 4b,c). It is found by comparing the POAO run with the LO21 run that the presence of subtropical zonal LSD significantly intensifies the atmospheric heat sources and moisture sinks over the South China Sea region (Fig. 4c). The influence of tropical continents and Asian and African mountains on $h_{Q_1}$ and $h_{Q_2}$ are not significant. In comparison with the LSD and orography, SST shows the most significant influence on $Q_1$ and $Q_2$ in the South China Sea region. The $Q_1$ ($Q_2$) increases from 34 (30) W m$^{-2}$ in the M run to 113 (83) W m$^{-2}$ in the M2 run. In the M2 run, the vertical profiles of $Q_1$ and $Q_2$ averaged over the South China Sea region are similar to those in Fig. 5c. This all suggests that the realistic SST distribution plays a dominant role relative to the LSDs and mountains in establishing cumulus convection over the South China Sea region. Considering the poor performance of CCM3 in simulating the East Asian subtropical rainfall, $Q_1$ and $Q_2$ in the East Asian region are not discussed here.

4. Possible mechanisms of the Asian summer monsoon response to LSD and orography

It is believed that the large-scale land–sea thermal contrast plays a dominant role in driving the Asian monsoon circulation. Although the importance of the tropical continents and the Tibetan Plateau in the Asian monsoon had been recognized by previous studies (e.g., Hahn and Manabe 1975; Dirmeyer 1998; Xu et al. 2001, 2002; Chou 2003; Liang et al. 2005, 2006), their influence relative to the large-scale meridional LSD still remains unclear. The results of section 3 indicate that the tropical subcontinental-scale zonal LSD and Asian mountains play a more important role than the large-scale meridional LSD between the Eurasian continent and the Indian Ocean in strengthening the MCII and SAMR. In this section, we will elucidate on the possible mechanisms of the Asian summer monsoon response to the tropical subcontinental-scale zonal LSD and the Asian mountains.

a. Tropical subcontinental-scale zonal LSD

The presence of the Indian subcontinent and Indochina peninsula significantly strengthens the summer monsoon circulation and rainfall over the continents and to their west (Fig. 3h). The main reason is likely as follows: In summer, the warm land surface tends to induce shallow circulation between the ocean and land. Then, maritime air with high moisture content converges into the low pressure system formed over the land, and cumulus convection and precipitation are formed. After the rain starts, the land–sea temperature contrast decreases but the low-tropospheric convergence continues because condensation in cumulus clouds drives the ascending motion (Webster 1987). The tropical east–west asymmetry heating induced by cumulus convection tends to induce a low-tropospheric low pressure to the northwest and increasing westerly winds to the west of the heating center (Gill 1980). The increasing westerly flow is usually accompanied by increasing precipitation due to the release of unstable energy and the plentiful moisture. Thus, the heating center appears to be shifted westward associated with the westward shift of convective
precipitation. As this process continues, the heat-induced circulation and precipitation will gradually shift toward the west.

Figure 6 shows longitude–time sections of the pentad-mean 850-hPa geopotential height anomaly averaged over 18°–22°N for the (a) POAO run, (b) AfrSAM run, (c) India run, and (d) SAsia run. Only the summer (June–September) sections are shown here for the last five years of simulation. The geopotential height anomaly is defined by the geopotential height minus the summer- and area-mean geopotential height averaged over 18°–22°N, 30°W–130°E. (e) As in (d) but for the daily geopotential height anomaly in the 1-yr NM run.

FIG. 6. Longitude–time sections of pentad-mean geopotential height anomaly (in 10 gpm) averaged over 18°–22°N for the (a) POAO run, (b) AfrSAM run, (c) India run, and (d) SAsia run. Only the summer (June–September) sections are shown here for the last five years of simulation. The geopotential height anomaly is defined by the geopotential height minus the summer- and area-mean geopotential height averaged over 18°–22°N, 30°W–130°E. (e) As in (d) but for the daily geopotential height anomaly in the 1-yr NM run.
persists for a relatively short period compared with that in the simulations with tropical continents (Figs. 6a–d). In the AfrSAm run, most low pressure systems and heavy rainfall are created over northern Africa west of \( 50^\circ \)E and show clear westward propagation (Fig. 6b). In the India run, however, low pressure systems appear over the Indian subcontinent (\( 70^\circ \)–\( 80^\circ \)E) and propagate westward in association with increasing rainfall, which indicates that the presence of the subcontinent is conducive to the formation and development of a low pressure system (Fig. 6c). When both the subcontinent and Indochina peninsula are introduced, more low pressure systems and heavy rainfall are simulated over the Indochina peninsula (\( 95^\circ \)–\( 110^\circ \)E) in the SAsia run compared to the India run (Figs. 6c,d).

To better show the time-dependent features of circulation and precipitation, a 1-yr CCM3 branch run with daily output was performed using the 13th year output from the NM run as the initial field. The daily \( 850 \)-hPa geopotential height (Fig. 6e), zonal wind, and precipitation for the NM run show clearer westward propagation than the pentad-mean ones owing to the higher temporal resolution. Clearly, the zonal heating contrast induced by the presence of tropical continents helps create the low pressure systems. The tropical oceans supply plentiful moisture to develop the low pressure system into cumulus convection, which in turn induces a strong response of the tropical monsoon circulation characterized by the westerly (easterly) anomaly to the west (east) of the deep convection. This can explain why the presence of the African continent slightly weakens the vertical shear of zonal wind over the South Asian region, whereas the presence of the Indian subcontinent and Indochina peninsula significantly strengthens it (Table 3).

b. Asian mountains

Figure 7 illustrates the annual cycle of the vertically integrated heating rate induced by the apparent heat source \( \langle Q_1 \rangle \) and apparent moisture sink \( \langle Q_2 \rangle \), as well as the sensible heat, latent heat of condensation, net radiation, and latent heat of evaporation in Eqs. (2) and (3) over the Tibetan Plateau region (elevation >3000 m) for the NM and MAsia runs. Correspondingly, Fig. 8 shows the pressure–time section of the difference of pentad-mean geopotential height averaged over \( 70^\circ \)–\( 100^\circ \)E between \( 25^\circ \) and \( 10^\circ \)N. In the NM run, sensible heat dominates the net vertically integrated heating rate over the Tibetan Plateau region (Fig. 7a). Whereas, release of the latent heat of condensation over the Tibetan Plateau increases dramatically and accounts for most of the net vertically integrated heating rate when the Asian mountains are introduced (Fig. 7b). It is also found from Fig. 7 that the sensible heating rate has a rapid increase in February, reaching its maximum in late March in the MAsia run, which is earlier than in the NM run. At the same time, the pressure gradient between \( 25^\circ \) and \( 10^\circ \)N is reversed over the longitudes between \( 70^\circ \) and \( 100^\circ \)E, which cannot be found in the NM run (Figs. 8a,b). This means the low pressure enveloping the Tibetan Plateau and its vicinity has emerged in spring, associated with the increased sensible heating over the Tibetan Plateau.
With the enhancement of the South Asian low, southwesterly winds are significantly strengthened over South Asia, which tends to evaporate more water vapor from the sea surface into the atmosphere. On the other hand, existence of the low pressure envelope helps to converge the moist southwesterly flow from the northern Indian Ocean onto the Tibetan Plateau. For the moist air parcels over the northern Indian Ocean, their free convection levels are often lower than the height of the Tibetan Plateau. Thus, the dynamical lifting of moist air induced by the plateau helps to trigger the cumulus convection. In turn, the convective release of latent heat from precipitation intensifies the low pressure envelope. This provides a positive feedback, thus amplifying the monsoon circulation. Although the heating rate induced by the release of latent heat from precipitation is 2–3 times larger than that induced by the sensible heat in the MAsia run, it does not mean that sensible heating over the plateau plays a minor role in the Asian monsoon. The plateau sensible heating helps to trigger cumulus convection over and around the plateau. In fact, when the plateau sensible heating was removed from the numerical model, the simulated summer monsoon circulation dramatically weakened (Wu et al. 2007).

5. Summary and discussion

We evaluate the relative contributions of various LSDs and mountains to the Asian monsoon extent and intensity by a number of numerical experiments. This study provides a more comprehensive understanding of the roles of various LSDs and mountains in the Asian monsoon system through AGCM simulations with prescribed SST. The major findings are summarized here with suggestions for further research.

In comparison with the mountains, the LSD plays a more fundamental role in determining the extent of Asian and African monsoons. In the model, the tropical monsoon does not occur unless the boreal continent extends south of 21°N. The tropical continents, such as the African continent and the Indian subcontinent and Indochina peninsula, play a crucial role in the tropical monsoon system. Inclusion of the African continent results in an African monsoon characterized by intense vertical shear of the zonal winds and strong seasonality of precipitation; however, it does not exert significant influence on the extent of the South Asian monsoon. When the Indian subcontinent and Indochina peninsula are introduced, the South Asian monsoon shows a significant southward extension associated with increasing seasonality of precipitation. Comparison of the MAsia with the NM run reveals that presence of the Asian mountains helps to extend the seasonally reversing wind northward to 40°N. However, the presence of Asian mountains does not show a clear influence on the extent of other monsoon regions. In addition, the SST distribution plays a dominant role in determining the extent of the Australian and South China Sea monsoon.

Release of the latent heat of condensation dramatically increases in the summer monsoon region when the tropical zonal LSD is introduced. In summer, the warm land surface is conducive to the formation of a low pressure perturbation. The abundant moisture over warm oceans makes it easier to develop the low pressure perturbation into cumulus convection. As a response to the cumulus convection, a westerly wind anomaly occurs.
to the west of the cumulus convection (Gill 1980). The vertical shear of the zonal wind to the south of the cumulus convection also appears to strengthened significantly when tropical zonal LSDs are introduced. It is worth noting that, although the size of the Indian subcontinent and Indochina peninsula is relatively small, their contribution to the vertical shear of zonal wind is larger than that of the large-scale meridional LSD between the Eurasian continent and the Indian Ocean over the South Asian region. The zonal difference of MCI distribution seems to play a similar role as the tropical zonal land–sea thermal contrast, which intensifies the monsoon circulation and rainfall.

The Asian mountains play an important role as the tropical subcontinental-scale zonal LSD in the formation of vertical shear of the zonal wind over the South Asian region. In spring, the rapidly increasing sensible heating on the Tibetan Plateau tends to induce a low pressure system enveloping the plateau and its vicinity, which helps convergence and uplifting of moist southwesty winds from the northern Indian Ocean onto the Tibetan Plateau. The ascending motion associated with the release of latent heat of condensation will further intensify the low-tropospheric South Asian low and the upper-tropospheric South Asian high. As a result, the monsoon circulation index MCI increases from 4.5 m s\(^{-1}\) in the NM run to 11.9 m s\(^{-1}\) in the MAasia run, whereas the South Asian summer monsoon rainfall increases from 8.6 mm day\(^{-1}\) in the NM run to 10.2 mm day\(^{-1}\) in the MAasia run. The tropical subcontinental-scale zonal LSD and the Asian mountains are equally important to the increase of MCI and more important than the large-scale meridional LSD between the Eurasian continent and the Indian Ocean.

In fact, the relative sensitivity of Asian monsoon to the LSD and orography appears to depend on how the monsoon indices are defined. For example, as we can see in Table 3, the MCI significantly increases from 0.7 m s\(^{-1}\) in the LO32 run to 8.8 m s\(^{-1}\) in the LO21 run; however, the South Asian summer rainfall significantly decreases from 8.2 mm day\(^{-1}\) in the LO32 run to 6.7 mm day\(^{-1}\) in the LO21 run. The increase of MCI largely results from intense sensible heating of the subtropical continent, not the release of latent heat of condensation. If the increase of MCI mainly results from cumulus convection, as in the SAasia run, the MCI and South Asian rainfall will both increase significantly because cumulus convection enhances the low-level South Asian low and upper-level South Asian high, which in turn will enhance the vertical shear of the zonal wind south of the cumulus convection region. In the real world, the Asian summer monsoon is characterized by cumulus convection, so the increase of vertical shear of the zonal wind is usually accompanied by increasing rainfall. Clearly, to a certain degree, different monsoon indices are related to different forcing: some forcing will result in a change of monsoon indices in the same direction, but other forcing will not. The important detail is which forcing is dominant. Therefore, a multi-index evaluation is helpful to better understand the role of various LSDs and mountains in the Asian monsoon.

In this study, we employ an AGCM with prescribed SST to explore the response of monsoons to various LSDs, mountains, and SST distributions. The prescribed zonal mean SST removes the information of present-day LSD and orography stored in the SST and isolates the zonal land–sea thermal contrast from the zonal difference of SST. Only the zonal heating contrast between land and ocean is retained; otherwise, the zonal land–sea contrast is mixed with the zonal difference of SST. This is the merit of using a prescribed zonal-mean SST. However, there are two sides to the story. An AGCM simulation with prescribed SST cannot reflect the effects of air–sea coupling. The coupled AGCM simulation is able to describe the effects of LSD and orography more reasonably. As we can see from the previous sections, a differential SST distribution exerts considerable influence on the Asian summer monsoon. It is likely that air–sea interaction modifies the contributions of LSD and orography to the Asian monsoon. Some studies also suggested that coupled air–sea processes are crucial in the simulation of monsoons (e.g., Kitoh, 2004; Wang et al. 2005; Zhou et al. 2008). Therefore, a fully coupled atmosphere–ocean–land model will enable us to further examine the relative contributions of various LSDs and mountains to the Asian monsoon. In addition, although the present horizontal resolution of the model can well represent the effect of subcontinent-scale LSDs, it is still too coarse to capture the effect of small-scale mountains such as the Ghats mountain range, which plays an important role in the observed South Asian summer monsoon precipitation (Xie et al. 2006). High-resolution simulations are also desired. These are potential directions for future studies.

Acknowledgments. The authors thank the anonymous reviewers for their valuable comments and suggestions. The reanalysis data was provided by NCEP/NCAR. The Climate Prediction Center Merged Analysis of Precipitation (CMAP) data was provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado (http://www.cdc.noaa.gov/). This research is supported jointly by the “National Basic Research Program of China (973 Program)” Project 2006CB400500, China Postdoctoral Science Foundation 20070410133, and the Foundation of Jiangsu Key Laboratory of Meteorological Disaster, KLME0704.
REFERENCES


Wu, T.-W., and Z.-A. Qian, 2003: The Relation between the Tibetan winter snow and the Asian summer monsoon and


