Influence of the Land Surface in the Asian Summer Monsoon: External Conditions versus Internal Feedbacks*

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ABSTRACT

The basic concept of land–sea temperature contrast and the strength of the Asian summer monsoon is investigated here by comparing the relative contributions of external conditions (including surface albedo) and internal feedbacks (involving soil moisture) in a number of atmospheric general circulation model (GCM) mean climate simulations and in a GCM sensitivity experiment. All models are run with the same long-term mean sea surface temperatures so that only land-surface conditions affect the land–sea temperature contrast. There is a surprising consistency among the various models such that stronger summer monsoons (defined as high area-averaged precipitation over south Asia) are associated with greater land–sea temperature contrast (i.e., higher land temperatures), lower sea level pressure over land, less snow cover, and greater soil moisture. In a sensitivity study with land albedos uniformly raised from 0.13 to 0.20 in one of the models, the winter–spring–summer sequence over southern Asia shows that there is a high sensitivity to the specified land albedos. Lower land albedos are associated with warmer land temperatures, greater land–sea temperature contrast, and a stronger Asian summer monsoon. There is also a positive feedback between soil moisture and precipitation (increased soil moisture provides a surface moisture source for further precipitation).

1. Introduction

The most fundamental concept of monsoon circulations is that of land–sea temperature contrast [e.g., see review by Webster (1987)]. As a landmass heats with the approach of summer, air at the surface is likewise warmed and begins to rise. This ascent draws in air at low levels in the atmosphere from the adjacent oceans to feed further convection that produces warming associated with latent heat release. With the proper orientation of land and sea, this circulation can become regionally extensive. Consequently, seasonal monsoon circulations become established over many areas of the Tropics (Ramage 1971) with links to weather and climate fluctuations in other parts of the globe (e.g., Kiladis and van Loon 1988; Lau 1992; Webster and Yang 1992). Variations of land–sea contrast have been identified as important in terms of the role of topography in producing such contrasts (e.g., Hahn and Manabe 1975; Meehl 1992). Accentuation of land–sea temperature contrast also appears to be crucial for stronger monsoon circulations associated with changes of solar and other external forcing factors on paleoclimatic timescales (e.g., Kutzbach and Guetter 1986; Prell and Kutzbach 1992; Kutzbach et al. 1993).

Even though land–sea temperature contrast has been shown to be linked to mean monsoon strength, there seem to be several factors that can contribute to that contrast. One category involves external conditions set up (or specified) prior to the monsoon season. Another involves internal feedbacks that occur during the monsoon season. These feedbacks have been studied in a number of contexts for various regions in general circulation model (GCM) experiments [see review by Garratt (1993)]. In this paper we will examine external conditions related to surface albedo and internal feedbacks involving soil moisture and study their relative contributions specifically with regard to land–sea temperature contrast and the strength of the south Asian monsoon in a number of GCMs.

Charney et al. (1977) and others have noted that higher land albedos (possibly due to land use change) can reduce seasonal rainfall by reducing the absorbed solar radiation at the surface. This can lead to cooler land temperatures, decreased land–sea temperature contrast, less rainfall, and a weak monsoon (Fig. 1, left side). A weak monsoon here is defined as below-normal area-averaged rainfall over south Asia. A weak monsoon is also typically characterized by reduced southwesterly inflow from the Arabian Sea and a weakened
tropical easterly jet over India, and vice versa for a strong monsoon (e.g., Krishnamurti et al. 1989, 1990; Palmer et al. 1992). Additionally, the idea of a reduction of land–sea temperature contrast resulting from greater albedo over south Asia from increased snow cover, with a consequential weakening of the Indian monsoon, was introduced in the last century (e.g., Blanford 1884) and has been studied using satellite-derived snow cover data (e.g., Hahn and Shukla 1976). This concept has been addressed more recently with GCM simulations that basically confirm the earlier results (e.g., Barnett et al. 1989; Yasunari et al. 1991). As snow melts and moistens the surface, the albedo effect gives way to a soil moisture feedback. Thus, snow-cover changes can contribute to both external conditions prior to the monsoon and internal soil moisture feedback during the monsoon. However, the connection between Tibetan snow cover and the south Asian monsoon is not always a clear one in GCM experiments (Zwiers 1993) and may have a variety of forcings from other regions (Chen 1994). For our purposes, we categorize the snow-cover albedo effect as an external condition that is strongly a consequence of the specification of other external conditions such as bare land albedo. Subsequent changes in soil moisture fall into the internal feedback category.

In addition to alterations of surface albedos as an external forcing condition that can affect land–sea temperature contrast prior to the monsoon season, internal feedbacks involving soil moisture can provide a significant influence on rainfall during the monsoon season. For example, Meehl and Washington (1988) note a soil moisture–precipitation feedback that can maintain positive anomalies of each through a summer season. This mechanism has been discussed in various contexts (e.g., Namias 1959; Walker and Rowntree 1977; Shukla and Mintz 1982; Yeh et al. 1984; Delworth and Manabe 1989; Simmons and Lynch 1992), including ones pertaining to intraseasonal (e.g., Webster 1983; Nanjundiah et al. 1992) and interannual (Yasunari and Seki 1992) fluctuations of Indian monsoon precipitation. In the context of the south Asian monsoon and land–sea temperature contrast, this feedback can seemingly be either positive or negative. For soil moisture feedback to be positive (far right side of Fig. 1), elevated soil moisture induces greater evaporation that provides an increased moisture source for enhanced precipitation, a strong monsoon, further increases of soil moisture, and so on. However, the feedback can also be negative (center of Fig. 1), as the increased evaporation from greater soil moisture amounts causes a cooler land surface, decreased land–sea temperature contrast, less precipitation, and a weak monsoon that then results in decreased soil moisture amounts.

Therefore, two competing effects exist with regard to internal soil moisture feedback—local enhancement of precipitation with higher levels of soil moisture, and a regional decrease of monsoon precipitation with the decrease of land–sea temperature contrast and weaker monsoon flow. In a study by Barnett et al. (1989), the negative feedback was most important during the summer monsoon season. In a later study by Yasunari et al. (1991), the large decrease of monsoon precipi-
tation noted by Barnett et al. was not as evident apparently because of positive soil moisture feedback. Yasunari et al. pointed out that whether soil moisture feedback was positive or negative was not clear and could have to do with the type of convective scheme in the model.

The purpose of this paper, therefore, is to examine the relationship between land–sea temperature contrast and monsoon strength in a number of GCM mean climate simulations and then explore the contributions to that contrast for external conditions involving land-surface albedo and internal feedbacks involving soil moisture in a sensitivity experiment. This study differs from earlier ones (e.g., Barnett et al. 1989; Yasunari et al. 1991) in that multiyear climatologies from a number of models with different mean monsoon characteristics are compared to address these issues.

Because large-scale, east–west circulations in the Tropics are related to regional precipitation characteristics (e.g., Meehl 1987), the global tropical simulations of several models are first shown to provide a context for the regional monsoon discussion. These models are chosen to have representative simulations of features that appear in a number of the other simulations. Then area averages are computed over the south Asian monsoon region and compared for all the models. Finally, the monthly evolution (January through August) of hydrological and surface energy balance components is presented for this same area in a subset of the models (again chosen to be representative) to examine the factors contributing to the land–sea contrast and monsoon strength. In general, the broad-scale regional Asian summer monsoon is the subject of this analysis. The rainfall over India itself, while certainly an important component of the south Asian summer monsoon regime, is not addressed explicitly. However, the rainfall over India is a part of the regional-scale, summer monsoon regime and, in many ways, is implied in the larger-scale, regional monsoon characteristics.

2. Experimental design

GCM experiments usually take two forms. First, a number of different model versions can be compared, all with their own formulations and systematic errors, to study how the models simulate some feature of interest. If consistent results emerge from such a comparison, insight can be gained concerning how a process or mechanism is represented in each of the different models.

A second technique involves running one model with different conditions (ideally changing only one thing at a time), and the “experiment” can be compared to the “control” integration to study what changes occurred in the simulated climate (often referred to as a sensitivity study). These changes in some process or simulated feature can then be better understood in terms of what was changed in the experimental integration.

In this paper both approaches are used. First, a number of models with different configurations are compared to determine whether there is a consistency among the models concerning land–sea contrast and the strength of the Asian monsoon. It is not the purpose of this paper to do an exhaustive intercomparison [this is being undertaken elsewhere and with a larger group of models; e.g., see World Climate Programme (1992)]. Instead, this comparison is done to determine whether there are any common features in the various simulations with regard to land–sea contrast and monsoon strength.

After having established the characteristics associated with land–sea surface temperature contrast in the model simulations, a sensitivity study is performed to address the role of factors that could contribute to setting up or maintaining that contrast. Then, land-surface albedo is changed in one of the models to test the specific effects on land–sea contrast and monsoon strength. By combining results from using both GCM experimental techniques, we attempt to gain insight into the relative roles of external forcing (associated with surface albedo) and of internal feedbacks (from soil moisture) in contributing to land–sea temperature contrast and monsoon strength.

It is important to keep in mind that these are long-term mean climate experiments, not forecast experiments. All the model results are from equilibrium multiyear seasonal cycle climate integrations with fixed climatological sea surface temperatures (SSTs). Multiyear averages are computed (Table 1) after a sufficient period of time (usually several years) such that the effect of initial conditions is negligible [for climate model integration strategies, see Washington and Parkinson (1986) or Trenberth (1992)]. By computing multiyear averages, the effects of intraseasonal transients (e.g., Chen et al. 1988; Gadgil and Srivivasan 1990) are largely removed. Studying changes in monsoon intensity in GCMs with fixed SSTs necessarily omits important feedbacks involving changes in SSTs (e.g., Clemens and Oglesby 1992). Interannual variability of the entire monsoon system (including SSTs) is the subject of a subsequent study using a global coupled ocean–atmosphere GCM.

3. The models

Table 1 is a summary of the various model versions and a few of their salient features. These particular models were chosen on the basis of common origins [National Center for Atmospheric Research Community Climate Model (NCAR CCM0) and Australian Bureau of Meteorology Research Centre (BMRC)] or common institutional development (NCAR CCM0 and CCM1). Thus, there are some similar elements in the various versions, but they are different enough to provide a range of monsoon simulation characteristics. The Monsoon Numerical Experimentation Group
Table 1. Summary of key features of GCM versions, all using Alexander and Mobley (1976) SSTs. Model abbreviations in first column are used in Fig. 9. CCM1 BATS uses a two-level, soil moisture scheme, and in the CCM0 albedo experiment land albedos are raised from 0.13 to 0.20.

<table>
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<th>Model</th>
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<td>9</td>
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<tr>
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<td>R15</td>
<td>12</td>
<td>Convective adjustment</td>
<td>BATS</td>
<td>3 yr</td>
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Sensitivity experiment

CCM0 albedo**          R15     9  Convective adjustment  Bucket  3 yr

* BMRC specifies snow cover over Tibet.
** Same version as CCM0 except with land albedo from 0.13 to 0.20.

(MONEG) is producing a set of model runs in forecast mode (i.e., individual seasons with specific initial conditions). These simulations are not appropriate for the types of mean climate analyses to be done in this paper. Another large set of models with some boundary-forcing features in common is being produced as part of the Atmospheric Model Intercomparison Project (AMIP). However, these models are being run with time-varying SSTs. For the purposes of this study (and to isolate the effects on land–sea contrast of external forcing conditions or internal feedbacks over land), it was desired that all models be integrated with the same climatological-specified SSTs in order to form consistent multiyear averages.

The NCAR CCM0 versions contain nine levels and an R15 horizontal resolution (rhomboidal 15 truncation yielding a grid of about 4.5° latitude by 7.5° longitude). One version uses simple convective adjustment and computed clouds, as described in Washington and Meehl (1984) and Meehl (1989), and the other includes the Albrecht hybrid mass flux convective scheme with clouds computed in the same manner (e.g., Meehl and Albrecht 1988, 1991). The NCAR CCM1 has 12 levels and is represented at two resolutions—R15 and T42—with computed clouds and is described by Williamson et al. (1987). The latter resolution yields a grid of about 2.8° latitude by 2.8° longitude. The BMRC GCM has nine levels and an R21 resolution (about 3.3° latitude by 5.6° longitude) with computed clouds and is described by Hart et al. (1990) and McAvaney et al. (1991). All of these models use the simple, bucket soil moisture formulation [a 15-cm soil moisture reservoir; see description in section 7 and in Meehl and Washington (1988)] with surface air temperature equivalent to the temperature in the lowest model layer (σ = 0.991 in all models). Somewhat different from the previous models in terms of surface hydrology is a version of CCM1 R15 with the Biosphere–Atmosphere Transfer Scheme (BATS) surface hydrology (Dickinson et al. 1986) with a two-layer soil moisture formulation and surface air temperature calculated at 1.5 m.

As discussed in the section on experimental design (section 2), a version of CCM0 (with convective adjustment) is run in a sensitivity experiment, with land albedos over the monsoon area raised from 0.13 (as specified in the standard CCM0) to 0.20. Presumably, one could imagine this as a test of monsoon sensitivity with changing land use that would alter the vegetation and thus the albedo. These albedos also clearly fall short of schemes that specify land albedo based on detailed land-surface types. However, the intent here is simply to test this model's sensitivity to the external forcing involved with different large-scale specified land albedos. Snow area (and thus snow albedo feedback) is not directly altered. However, as will be seen shortly, snow cover is directly affected by the role of land albedo in surface energy balance and resulting land temperatures.

All model versions use the same specified monthly varying SSTs from Alexander and Mobley (1976). Therefore, only land-surface conditions can affect the land–sea temperature contrast. Atmospheric winds simulated by several model versions are compared to mean winds from the European Centre for Medium-Range Weather Forecasts (ECMWF) analyses from 1979 to 1988.

4. Global tropical simulation

Figure 2 depicts Northern Hemisphere summer mean precipitation (June–September, multiyear averages) simulated by several of the models along with observed precipitation from Jaeger (1976). Although there are other datasets of observed precipitation, the comparisons to follow show greater-magnitude model-to-model discrepancies than differences between observed datasets. The tropical simulations in these model versions are representative of other models in this study, as noted where appropriate in the following discussion.

The model with the simulation that most closely resembles the observed distribution of seasonal precip-
Fig. 2. Seasonal (June–September) mean precipitation (mm day$^{-1}$) for (a) observations from Jaeger (1976), (b) CCM0 R15, (c) CCM1 R15, and (d) CCM1 T42. The first contour plotted in all four frames is 2 mm day$^{-1}$, with a successive contour interval of 2 mm day$^{-1}$. Stippling indicates areas greater than 4 mm day$^{-1}$. 
itation in the south Asian summer monsoon precipitation is the NCAR CCM0 (Fig. 2b). In this model, as well as the version with the Albrecht hybrid convection scheme (not shown), extensive precipitation (values in excess of 4 mm day\(^{-1}\)) covers most of India, extends into Bangladesh and Burma, and is similar to the observed distribution (compare CCM0 in Fig. 2b to the observed values in Fig. 2a). Precipitation maxima up to 12 mm day\(^{-1}\) are present in western India in the CCM0 compared to about 10 mm day\(^{-1}\) in the Jaeger data. There are peak values in the model of 18 mm day\(^{-1}\) shifted somewhat north over Bangladesh, while 18 mm day\(^{-1}\) maximum values in the observations lie somewhat to the southeast over Burma. This model also represents the rainfall maximum of 6 mm day\(^{-1}\) over northern South America. The CCM0 simulates other major areas of tropical precipitation but with greater-than-observed precipitation over tropical eastern Africa and near New Guinea (about twice the observed in both locations) and less-than-observed values by about 40% over Central America.

Anomalously low precipitation over India (about 2 mm day\(^{-1}\) or less) with high precipitation just east of the Philippines (about twice the observed) characterizes the models considered here that simulate a weaker-than-observed Indian summer monsoon—CCM1 at both resolutions (Figs. 2c and 2d) as well as BMRC (not shown). The location of the rainfall maximum over west Africa is well represented but with values half to two-thirds of the observed values. The greater-than-observed precipitation in eastern Africa similar to CCM0 is probably related to poorly resolved topography there. Location of rainfall maxima over northern South America is also well simulated but with about twice the observed values in the CCM1 R15 and considerably more in the T42 version. Excessive rainfall also appears over Central America at both resolutions. As noted above, the general characteristics of the rainfall pattern in the CCM1 between R15 and T42 are similar, but the T42 shows more regional detail and a better-resolved intertropical convergence zone (ITCZ) in the eastern Pacific. The main differences between the CCM0 and CCM1 precipitation simulations are over Central America, India, and the western Pacific near the Philippines.

These features of the precipitation field are reflected in the soil moisture (Fig. 3; recall that field capacity for the bucket soil moisture formulations is 15 cm) and the low-level (850 mb) wind fields (Fig. 4) in these models. The CCM0 has high soil moisture (in excess of 4 cm) extending over most of India and soil moisture maxima of about 12 cm over the other tropical continents, corresponding to the precipitation maxima of Fig. 2. The CCM0 low-level wind field is comparable to the observed but with somewhat larger-magnitude winds. In particular, strong westerlies in excess of 10 m s\(^{-1}\) across continental Africa just north of the equator are associated with the greater-than-observed precipitation in eastern Africa, and the easterlies are too strong in the tropical Pacific (by about a factor of 3) and across South America (by about a factor of 2). The CCM1 (at both resolutions), as noted above, has less precipitation in the South Asian area, especially over India, and greater-than-observed precipitation east of the Philippines in the western tropical Pacific. These precipitation characteristics are associated with weak (by about 30%) and southward-shifted southwesterlies over the Arabian Sea, anomalously strong westerlies (by a factor of 2) extending over Southeast Asia, and an anomalous cyclonic circulation east of the Philippines. The BMRC model with a precipitation simulation similar to the CCM1 also exhibits very similar low-level wind flow characteristics (not shown).

Figure 5 depicts upper-level winds at 200 mb for the models and the ECMWF analyses. For the CCM0 (Fig. 5b), strong tropical easterlies (about 20 m s\(^{-1}\)) associated with the vigorous precipitation over India and Africa noted in Fig. 2 extend from Southeast Asia across the Indian Ocean and Africa to South America. For the CCM1 versions at both resolutions (Figs. 5c and 5d), the easterlies are confined to the northern Tropics and are far less extensive (associated with decreased precipitation over the Indian region) than the ECMWF analyses (Fig. 5a).

The results of the global tropical simulations point to east–west linkages that are associated with certain aspects of the precipitation over the South Asian monsoon region. Systematic errors in the simulation in other regions could affect precipitation over south Asia (e.g., in CCM1 the anomalous precipitation over the Philippine region is associated with deficient monsoon precipitation over India). For the remainder of this paper, however, we focus on the land-surface conditions over south Asia in association with the precipitation simulation there to determine what consistency, if any, exists in regional forcing of the monsoon circulation.

5. Land-surface conditions and the Asian summer monsoon

The relationship between land-surface conditions and the monsoon is explored in terms of area-averaged surface air temperature, sea level pressure (SLP), soil moisture, snow cover, and precipitation over the Asian subcontinent in June. This is the time of monsoon onset and the culmination of conditions in winter and spring over land that contribute to subsequent monsoon strength. June is considered representative of those conditions to characterize the state of the summer monsoon season in the model simulations (e.g., see Barnett et al. 1989). This may not, in fact, be the case in every observed monsoon, but June is appropriate for this study to represent first-order aspects of the land-surface conditions and monsoon precipitation for long-term mean monsoon climate. It will be shown in sec-
tion 6 that multiyear averages of monsoon seasonal characteristics in the models are well represented in the multiyear June averages.

The quantities under consideration are averaged for land grid points only for the area 5°–40°N, 60°–100°E (Fig. 6), which includes the major precipitation maxima for the south Asian monsoon (Fig. 2a). Previous studies with increased snowfall (e.g., Barnett et al. 1989; Yasunari et al. 1991) considered a larger domain for land effects that included the entire Eurasian continent. Here, the smaller area involved with the actual monsoon precipitation itself is considered, but the results are generally representative of conditions over the larger Asian area.

All area-averaged quantities are plotted as a function of area-averaged precipitation and are shown in Fig. 7. BMRC snow cover over the Tibetan plateau is partly a function of prescribed land-surface albedo. Consequently, snow cover and soil moisture (both affected by snow cover) are not plotted for BMRC in Figs. 7c and 7d. The CCM1 version with BATS includes a two-layer soil moisture scheme. Therefore, this model is in
VECTOR WIND 850 MB (JUN-SEP)

ECMWF

CCMO R15

CCMI R15

CCMI T42

Fig. 4. Seasonal (June-September) vector winds at 850 mb for (a) ECMWF analyses (1979-88), (b) CCM0 R15, (c) CCM1 R15, and (d) CCM1 T42 (vectors plotted at every other grid point). Scaling vectors at lower right of each frame.
Fig. 5. Seasonal (June–September) vector winds at 200 mb for (a) ECMWF analyses (1979–88), (b) CCM0 R15, (c) CCM1 R15, and (d) CCM1 T42 (vectors plotted at every other grid point). Scaling vectors at lower right of each frame.
a somewhat different class compared to the bucket soil moisture schemes, and values for soil moisture from this model are not included in Fig. 7d. More detailed analysis of the monsoon simulation in this particular model version is the subject of a separate study.

Even though these model integrations are mean climate simulations with fixed SSTs, the soil moisture amounts and surface conditions vary somewhat from year to year. Typical interannual standard deviations of the area-averaged quantities plotted in Fig. 10 are considerably smaller than the differences of mean quantities between models. Magnitudes of standard deviations are also quite similar between the models.

For example, typical standard deviations for area-averaged precipitation are about several tenths of a millimeter per day, for SLP on the order of one millibar, around several tenths of a degree for surface temperature, and about several tenths of a centimeter for soil moisture. Mean differences of area-averaged quantities between the models for these parameters usually exceed at least two times these standard deviations.

Figure 7 shows a surprising consistency in models having quite different characteristics, with a stronger summer monsoon (indicated by greater amounts of area-averaged precipitation) associated with higher temperature, lower SLP, less snow cover, and higher soil moisture over the south Asian monsoon region. The addition of a computed surface process scheme (BATS) in CCM1 increases the surface air temperature (calculated by BATS), decreases SLP, and increases monsoon precipitation. Addition of the Albrecht mass flux convection scheme in CCM0 does not substantially affect the strength of the monsoon.

Results from the surface albedo sensitivity experiment (discussed in more detail in section 6) are also shown for the version of CCM0 where land-surface albedos are raised from 0.13 to 0.20. The resulting monsoon simulation is weaker than the standard CCM0 with decreased precipitation, lower surface air temperature, higher SLP, and less soil moisture.

In Fig. 7a, two estimates from observational datasets that include both gridded precipitation and surface air temperature (Shea 1986; Legates and Wilmott 1990) show a similar relationship to that seen in the models with higher temperature associated with higher monsoon precipitation. Both observational datasets are shifted toward higher values of temperature and somewhat lower values of precipitation compared to the models.

6. Sensitivity to land albedo

Since land albedos are specified in a nonuniform manner in the various model versions, it is difficult to determine exactly what role this important external forcing plays in the manifestation of the monsoon in the models. Therefore, an experiment is performed with the CCM0 to test the sensitivity of the monsoon simulation to a change of land albedo. In this experiment (called CCM0 albedo), land albedos are raised from 0.13 to 0.20. Results from this experiment are computed for January through August and compared to the standard CCM0, as well as to the CCM1 at T42. The CCM1 has an area-averaged land albedo over the south Asian region (Fig. 6) of about 0.18 but, as noted below, has a much higher effective total surface albedo owing to greater snow cover than in either of the CCM0 versions. The results for June are those shown in Fig. 7. Since the effects of a change in external condition in the months prior to the monsoon could affect the precipitation during the monsoon season, results in Figs. 8–12 are shown for the January through August monthly evolution.

As noted earlier in Figs. 2–5, the standard CCM0 was judged to produce a better monsoon simulation in terms of distribution of monsoon rainfall, soil moisture, and winds compared to the other model versions. When the specified land albedo is raised in this model, the land temperature decreases in all months by about 2°C (Fig. 8), thereby weakening the land–sea temperature contrast. Monsoon precipitation (Fig. 9) decreases accordingly by about 1 mm day⁻¹ or less in the months prior to the monsoon, and near 2 mm day⁻¹ (or about 25%) during the June–August monsoon season. The CCM1 simulates about half the area-averaged monsoon precipitation of CCM0 (Fig. 9), and the temperature over land in every month is lower than both of the CCM0 versions (Fig. 8) by several degrees (signifying decreased land–sea temperature contrast). Associated with these temperature differences, SLP in Fig. 8 is lower in the CCM0 standard version with the stronger monsoon than in the CCM0 albedo experiment by about 1 to 2 mb prior to the monsoon, and by around 4 mb during the monsoon season. This is representative of stronger low-level inflow from ocean to land. The CCM1 version has higher SLP than the CCM0 version during the premonsoon and monsoon onset period from March to July by about 3 to 5 mb.

Consistent with the precipitation results, Fig. 9 also shows that the models with greater precipitation have comparatively higher evaporation. (The negative values of evaporation denote a contribution to decreasing surface moisture; “greater values” here refer to larger
negative values in Fig. 9.) By raising the land albedos, the CCM0 shows decreased evaporation on the order of 1 mm day$^{-1}$ or around 20% during the monsoon season. Because of a combination of lower surface temperature (Fig. 8) and less precipitation (Fig. 9, top), the CCM1 has less than half the evaporation than in the standard CCM0 (Fig. 9, bottom).

The higher the rainfall in the models, the greater the soil moisture (Fig. 10). Specifically, the standard CCM0 has more soil moisture (by about 20%-80%) throughout the period leading up to and including the monsoon (except during April) and is associated with greater precipitation and evaporation in Fig. 9 compared to both the CCM0 with increased land albedos and the CCM1. The models with greater precipitation and evaporation also have increased cloudiness (since the focus of this paper is on surface processes, cloud amounts are not shown here but will be mentioned in the context of rainfall since precipitation changes are almost directly proportional to cloud changes and can be considered as a suitable analog). This has a bearing on the radiative components to be discussed shortly with regard to Fig. 11.

The warmer temperatures in both versions of the CCM0 (Fig. 8) are associated with a total lack of snow cover in this region of south Asia, whereas the standard CCM1 maintains considerable snow cover (Fig. 10, given in units of liquid water equivalent) throughout
the period leading up to and including the summer monsoon. Not surprisingly, the standard CCM1 with the greatest snow cover has a much higher net surface albedo (land and snow as mentioned earlier) and the lowest surface air temperatures compared to the other model versions (Fig. 8).

Evaporation is stronger in the models with greater precipitation (Fig. 9). As noted in Fig. 1 this could act as a negative feedback by cooling the land surface and decreasing land–sea temperature contrast. However, the cooling effect of the moist land surface is offset by other factors that contribute to the maintenance of warmer land surfaces in the models in spite of greater soil wetness, thus causing soil moisture feedback to be positive. Figures 11 and 12 depict components of the surface energy balance for the three models. In comparing the standard CCM0 to the CCM0 albedo case in Fig. 11, the increased land albedos in the latter model are associated with about 9–12 W m\(^{-2}\) less net solar radiation absorbed at the surface in all months. That is, more solar radiation is reflected by the brighter surface even though there is less cloudiness and more solar radiation is incident at the surface (not shown). There is also a slight increase (several W m\(^{-2}\)) of net longwave radiation at the surface (i.e., more longwave radiation emitted upward). This latter effect is brought about partly by the decrease of clouds (mentioned above) that is associated with the reduced precipitation; that is, less cloudiness means less downward longwave radiation and greater net upward longwave radiation from the surface. These effects cool the surface and contribute to decreased land–sea temperature contrast and less monsoon precipitation in the CCM0 albedo experiment compared to the standard CCM0 (Figs. 8 and 9).

Comparing the standard CCM1 with the CCM0 versions, the CCM1 has less absorbed solar radiation at the surface compared to the standard CCM0 in all months, indicative of the higher land albedos and greater snow cover. The absorbed solar radiation in June and July in the CCM1 exceeds the totals for the CCM0 albedo case during those months. The average land albedos are somewhat less in the CCM1 than in the CCM0 albedo case (about 0.18 versus 0.20), so when the snow cover in June retreats to only a small part of the monsoon area in the CCM1 (not shown), the net surface albedo is marginally less (by about 5%), resulting in relatively greater values of absorbed solar radiation during those months.

There is an increase of net longwave radiation in the CCM1 of about 30 W m\(^{-2}\) in all months (decreased precipitation, decreased clouds, less downward longwave radiation, and increased longwave radiation emitted upward). In the standard CCM0 the greater absorbed solar radiation and decreased net longwave radiation emitted upward overcome the effects of the increased evaporation from the wetter land surface and lead to warmer land temperatures, greater land–sea temperature contrast, and increased monsoon precipitation. Soil moisture

![Diagram](image_url)

**Fig. 8.** Time series of monthly mean area averages (for area in Fig. 6), January through August, for surface air temperature (K, top) and SLP (mb, bottom). Standard CCM0 is solid line, standard CCM1 at T42 is dashed line, and CCM0 with the increased land albedos is long dash–dot line.

![Diagram](image_url)

**Fig. 9.** Time series of monthly mean area averages (for the area in Fig. 6), January through August, for precipitation (mm day\(^{-1}\), top) and evaporation (mm day\(^{-1}\), bottom). Negative units indicate evaporation is removing moisture from surface. Standard CCM0 is solid line, standard CCM1 at T42 is dashed line, and CCM0 with increased land albedos is long dash–dot line.
feedback then contributes positively as an increased moisture source at the surface for further rainfall. Additionally, the CCM0 albedo experiment shows that precipitation over land decreases by about 30%–40% while evaporation decreases only by about 25% during the monsoon months. This means that there is an increase of moisture flux convergence from the surrounding ocean regions as the monsoon inflow at low levels intensifies. This also contributes as a moisture source for greater monsoon precipitation (Yasunari et al. 1991; Kitoh 1992).

Figure 12 shows the sensible and latent heat flux area averages. From May through the monsoon season, the standard CCM0 has less sensible heat flux (about 5–10 W m⁻²) associated with the warmer air temperature at the surface compared to the CCM0 albedo experiment. The CCM1 has increases of 25–50 W m⁻² compared to the standard CCM0 during that same period, mainly owing to the much colder surface air temperatures. The standard CCM0 has increased latent heat flux values (compared to the CCM0 albedo version) of about 10 W m⁻² prior to May, and near 20 W m⁻² from May through August. The CCM1 has considerable decreases compared to both CCM0 versions as implied by the evaporation results in Fig. 9.

To summarize, factors in the models that contribute to greater land–sea temperature contrast and a stronger mean Asian summer monsoon with greater precipitation (standard CCM0 compared to the CCM0 albedo experiment and the CCM1) generally show 1) greater precipitation and evaporation, 2) higher air temperature at the surface, 3) lower SLP, 4) greater soil moisture, 5) less snow cover, 6) less sensible heat flux during the monsoon (and before as well in models without snow cover in the region) associated with the warmer air temperatures at the surface, 7) increased latent heat flux from the warmer moister surface, 8) increased net solar radiation absorbed at the surface (due in large part to decreased surface albedo and/or less snow cover), and 9) decreased net longwave radiation (mainly due to increased cloud cover associated with the increased precipitation). Therefore, the external conditions from surface albedo (specifically land-surface albedo in the CCM0 versions) are associated with positive internal soil moisture feedback (i.e., a moisture source for reinforcing monsoon rainfall).
7. Discussion

External conditions in the models (such as those associated with specified land albedos) can determine land–sea surface temperature contrast and can affect the mean monsoon strength and be dominant over the negative feedbacks from internal processes involving soil moisture feedback. Consequently, internal soil moisture feedback is positive in all the model simulations and contributes as a surface moisture source for increased monsoon rainfall.

Two previous model sensitivity experiments bear commenting on in the context of the present results. Barnett et al. (1989) and Yasunari et al. (1991) took steps to artificially increase snow cover over Asia in their model simulations to study the impact on the south Asian monsoon. Both groups found that the increased snow cover resulted in a cooler land surface, decreased land–sea temperature contrast, and less monsoon rainfall. However, details of the mechanisms differed somewhat between the two experiments. Barnett et al. (1989) found [as did Yasunari et al. (1991)] that increased snow cover in spring reflected more incoming solar radiation by means of increased surface albedo, and melting snow increased the soil moisture over land. The greater soil moisture was associated with increased evaporation, which kept the surface cool throughout the monsoon season, thus contributing to the weaker monsoon. Yasunari et al. found roughly the same effects in spring, but in summer the increased soil wetness provided a moisture source for additional convection and precipitation and associated heating in the midtroposphere. Consequently, the impact of increased snow cover in spring was blunted somewhat during the monsoon season. Monsoon rainfall decreased in total, but the effects were not as dramatic as in Barnett et al. Yasunari et al. also suggested a possible dependence of the soil moisture feedback on the convective scheme due to the tropospheric heating effects associated with the increased convection.

Results presented in this paper point out 1) the importance of soil moisture as a positive feedback (the far right side of Fig. 1), 2) the sensitivity of the models to specified land-surface albedos (the left side of Fig. 1), and 3) the implications concerning the possible relative role of external forcing and internal feedbacks for mean monsoon conditions compared to interannual variability. Even though none of the experiments discussed in this paper addressed the snow cover feedback directly, in the models with virtually no snow cover over Tibet (CCM0 versions), the land surface is warmer and wetter. Although the surface is cooled by greater evaporation, this effect is overcome by the positive feedback of lower specified land albedo and increased absorbed solar radiation at the surface, thus contributing to greater land–sea temperature contrast and stronger monsoon flow. Additionally, the moisture source over land contributes to greater precipitation.

In the versions with snow cover over the south Asian region, the standard CCM1 maintains snow cover before and during the monsoon season. It is associated with the coldest land temperatures, less soil moisture, and lowest monsoon precipitation of any of the models considered. This result agrees with the Barnett et al. and Yasunari et al. experiments and suggests that the contributions of snow cover to the role of the albedo effect in setting up regional-scale surface land–sea temperature contrast prior to the monsoon season is very important in terms of subsequent monsoon rainfall. However, soil moisture feedback is positive in all the present mean monsoon simulations, as opposed to the suggestions of Barnett et al. and Yasunari et al.

The CCM1 with BATS has more monsoon precipitation than the standard CCM1. This precipitation is still excessive over the Tibetan Plateau and in the western Pacific near the Philippines (as in the standard CCM1 in Fig. 2), but precipitation of
greater than 4 mm day$^{-1}$ extends over India (not shown). However, this larger precipitation total for CCM1 BATS is not indicative of a particularly improved spatial distribution of monsoon rainfall. Therefore, by changing details of the physical parameterizations in that model, some aspects of monsoon precipitation can be improved, but the overall pattern of monsoon precipitation remains roughly the same (e.g., see Rao et al. 1991).

Results presented here show the importance of positive soil moisture feedback for monsoon precipitation with implications for interannual variability of monsoon precipitation. The results suggest that (in an interannually varying interactive system, as opposed to mean conditions studied in the present paper), if a strong regional monsoon circulation is initiated early in the summer (e.g., through enhanced land–sea temperature contrast in late spring), the greater precipitation that falls as a consequence provides surface moisture to contribute to sustaining rainfall through the monsoon season by means of the soil moisture feedback mechanism. This could occur even as the surface land–sea temperature contrast is reduced owing to progressively cooler land temperatures, as the monsoon season proceeds mainly by means of increased latent heat flux from the wetter surfaces. Meanwhile, the midtropospheric temperature contrast is maintained by means of latent heat release from the increased convection to reinforce the positive feedback at the surface. These processes in an interannually varying system are being studied in a global coupled ocean–atmosphere GCM and will be the subject of a subsequent paper.

8. Conclusions

Simulation of the Asian summer monsoon with a number of different atmospheric GCMs, all run with the same observed SSTs, demonstrates that the fundamental concept of land–sea temperature contrast is largely determined by external forcing conditions, in particular land-surface albedo, that contribute to monsoon strength in the models. That is, a strong monsoon with heavy precipitation in June is associated with high temperature over the southern Asian land area, low SLP, low snow cover, relatively low specified land albedos, and high soil moisture. The addition of a land-surface process scheme (BATS) has the net effect of raising surface air temperature and intensifying monsoon precipitation. The addition of a mass-flux convective scheme does not affect monsoon strength substantially.

A sensitivity experiment performed with one of the models (with land albedo raised uniformly) shows decreased land–sea surface temperature contrast and a weakened monsoon. A comparison of this sensitivity experiment with the control run and another model version with a weak monsoon shows similar conditions in the months prior to and during the monsoon season. The stronger the simulated monsoon during this period, 1) the greater the surface air temperature, 2) the higher the evaporation, 3) the lower the SLP, 4) the greater the soil moisture, 5) the less the snow cover, 6) the less the sensible heat flux, 7) the greater the latent heat flux, 8) the greater the net solar radiation absorbed at the surface, and 9) the less the net longwave radiation emitted from the surface. Therefore, the effects of the external conditions involving surface albedo cause the soil moisture feedback to act in a positive sense (i.e., increased soil moisture is a moisture source for further rainfall in the context of enhanced land–sea temperature contrast from the altered surface albedos).

Soil moisture feedback is positive in all the model climate simulations with strong monsoons. The implication for mean monsoon conditions is that external conditions set up the large-scale land–sea surface temperature contrast that determines relative monsoon strength, and soil moisture acts as a positive feedback in a strong monsoon. Conversely, in the models with weak monsoons, decreased precipitation is associated with a dry land surface that contributes to further weakening of the monsoon.

In an interannually varying context (as opposed to mean monsoon conditions), a strong regional monsoon circulation could be initiated early in the monsoon season owing to large land–sea temperature contrast at the surface in late spring with contributions from a number of factors that could include decreased snow cover, warm air advection over continental Asia, positive surface energy balance, etc. The resulting monsoon rainfall could then receive a substantial contribution from positive soil moisture feedback in spite of falling land temperatures (and decreased land–sea temperature contrast) due mainly to increased latent heat flux from the wetter land surface. These processes are being explored in more detail in a subsequent study involving a global coupled ocean–atmosphere GCM.

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