Modelling the Asian summer monsoon rainfall and Eurasian winter/spring snow mass

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SUMMARY

The interannual variation of the south Asian summer monsoon is analysed based on sets of present day climate simulations using the UK Universities' Global Atmospheric Modelling Programme (UGAMP) general-circulation model with different land surface parametrization schemes, different horizontal resolutions, and sea surface temperature variations. Generally, a negative relationship is found between south Eurasian winter/spring snow mass and the amount of summer monsoon rainfall over India in all simulations. However, the significance of this relationship is dependent on the land surface parametrization scheme. The simulations using the no-flux boundary condition at the bottom of a three-layer soil model give a strong negative correlation. This inverse relationship is the strongest over north India and the foothills of the Himalayas and is statistically significant. The model results are in good agreement with observational studies. Composite analyses suggest that less winter/spring snowfall over south Eurasia is associated with a strong Indian summer monsoon, characterized by strong southwesterlies over the Arabian Sea in the lower troposphere in the June–August season and heavy precipitation in early summer over north India and the foothills of the Himalayas. In contrast, heavy winter/spring snowfall delays the onset of the Indian summer monsoon through the feedback of snowmelt, soil moisture and evaporation processes, and is associated with weak summer precipitation over the two regions.

Sensitivity studies confirm that the snow mass–Indian monsoon relationship identified in the simulation at T42 is also robust in the simulation at T31. However, a negative relationship does not exist in the simulation at T21, indicating the importance of horizontal resolution in maintaining the snow–monsoon relationship in the UGAMP model.

KEYWORDS: Asian summer monsoon General-circulation model Interannual variability Land surface parametrization

1. INTRODUCTION

The Asian summer monsoon is a dominant component of tropical climate variability and its influence extends to many regions remote from south-east Asia. The monsoon circulation is primarily driven by differential land–ocean heating and subsequently the release of latent heat by condensation. The seasonal variation of the solar insolation produces heating of the African and Asian continents during the northern hemisphere summer and cooling during the winter. Differential sensible heating results in the seasonal formation of low atmospheric pressure over the Asian and African continents and high atmospheric pressure over the surrounding oceans. The release of latent heat over southern Asia and northern Africa due to monsoon precipitation results in intensification of the low atmospheric pressure and enhances the monsoon circulation. The significance of the Tibetan Plateau as an elevated heat source for the onset of the Asian summer monsoon has been discussed by many authors (e.g. Hahn and Manabe 1975; Murakami 1987; Kutzbach et al. 1989, 1993; Yanai et al. 1992; Prell and Kutzbach 1992; Li and Yanai 1996). The eastern African highland also plays an important role (Rodwell and Hoskins 1995). Flohn (1957) suggested that the seasonal heating of the elevated surface of the Tibetan Plateau and consequent reversal of the meridional temperature and pressure gradients trigger the large-scale change of the general circulation over East Asia and the monsoon burst over the Indian subcontinent. Shukla and Fennessey (1994) pointed out that the annual cycles of solar forcing and sea surface temperature (SST) are important in the establishment of the Asian summer monsoon circulation and rainfall over India and adjacent regions. Changes in land–ocean temperature contrast also appear to be crucial for changing the monsoon cir-

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2567
calculation associated with change in the seasonal variation of solar forcing on paleoclimate scales (e.g. Kutzbach et al. 1989, 1993; Prell and Kutzbach 1992; Dong et al. 1996).

The Asian and Indian monsoons exhibit considerable interannual variability. This is clearly illustrated by year-to-year variations of the Indian monsoon rainfall (e.g. Sontakke et al. 1993). Various mechanisms have been proposed to explain this variability. These include SST anomalies in the tropical Pacific Ocean, Indian Ocean and the Arabian Sea (e.g. Shukla 1975; Washington et al. 1977; Fu and Fletcher 1985), Eurasian snow cover in the preceding winter/spring (e.g. Blanford 1884; Hahn and Shukla 1976; Dey and Bhanu Kumar 1983; Dickson 1984), and soil moisture (Shukla and Mintz 1982; Webster 1983).

The interannual variability of the Asian summer monsoon has been linked with the El Niño Southern Oscillation (ENSO) (e.g. Rasmusson and Carpenter 1983; Mooley and Shukla 1987; Webster and Yang 1992; Ju and Slingo 1995). Fu and Fletcher (1985) showed that the interannual variability of the Indian monsoon rainfall was highly correlated with that of the thermal contrast between the Tibetan Plateau and the equatorial Pacific.

Observational studies have suggested a significant negative correlation between the Indian monsoon rainfall and the Himalayan or south Eurasian winter snow cover (e.g. Blanford 1884; Hahn and Shukla 1976; Dey and Bhanu Kumar 1983; Dickson, 1984). This relationship is consistent with a suggestion by Charney and Shukla (1981) that the slowly varying surface boundary conditions play a role in the interannual variation of the Indian monsoon. Shukla (1987) hypothesized that excessive snowfall during the previous winter and spring seasons can delay the build-up of the monsoonal temperature gradient because part of the solar energy will be reflected and part will be utilized for melting the snow or for evaporating soil moisture. Some of the resulting meltwater is stored in the form of soil moisture, and this water continues to inhibit the surface sensible heating through the absorption of the latent heat of vaporization. Thus it is hypothesized that the Tibetan snow mass influences the strength of the Asian summer monsoon by delaying the time at which significant surface sensible heating can occur and by constraining the amount of heating that can take place. Recent observational studies using updated data and longer time series by Parthasarathy and Yang (1995) and Sankar-Rao et al. (1996) confirm the inverse relationship between the Eurasian winter snow cover and Indian summer monsoon precipitation. It is also suggested that the relationship becomes stronger in the partial correlation excluding the El Niño years.

There is also some indirect theoretical support for the snow–monsoon relationship. Yeh et al. (1983) found that removal of the snow cover can affect the overlying model atmosphere in a number of ways. Removal of the snow changes the surface albedo and increases absorption of incoming solar radiation. They also found that snow melt could affect soil moisture and subsequent evaporation. These changes had an impact on the temperature of the atmosphere from the surface to the upper troposphere. Meehl (1994) investigated the strength of the Asian summer monsoon by comparing the relative contributions of external conditions and internal feedbacks in a number of general-circulation model (GCM) mean climate simulations. All simulations used the same SSTs so that only the land surface condition affects the land–ocean temperature contrast. Results suggest that greater precipitation in June is associated with greater land–ocean temperature contrast, less snow cover, and greater soil moisture over the south Asian monsoon region.

GCMs have been used to investigate the physical mechanisms responsible for the relationship between Eurasian snow cover and the Asian summer monsoon. Barnett et al. (1989) described an experiment that was designed to test the response of the Asian summer monsoon to changes in snow cover and snowfall rate over Eurasia. They found that the prescribed albedo changes did not have a significant impact on the simulated Asian summer monsoon. However, prescribed snow accumulation-rate changes did have sig-
significant effects, with weaker monsoon precipitation during heavy snowfall simulations. Yasunari et al. (1991) did experiments to test the response of the simulated climate to prescribed changes in the thickness of the Eurasian snowpack in March. They found that the prescribed changes had a significant effect on the strength of the hydrological cycle in midlatitudes during the subsequent summer. They also reported a weaker Asian summer monsoon in response to an enhanced snowpack. Vernekar et al. (1995) reported a similar sensitivity that the anomalous greater spring Eurasian snow depth is linked to weak rainfall in the following summer over India. More-recent experiments performed by Douville and Royer (1996) using the Météo-France GCM give a similar response. However, Zwiers (1993) only found a weak relationship between the Eurasian snow cover and the Asian summer monsoon in a simulated climatology derived from two long integrations of the Canadian Climate Centre GCM. It was therefore suggested that the physical mechanism that is thought to connect the Tibetan snow with the Asian monsoon might not be properly simulated in the model. In particular, it was shown that the land surface scheme (one-layer model) does not store meltwater effectively. Consequently, the effect of the snowpack variations may be reduced to that of the corresponding albedo variation alone.

Most known empirical evidence is between the Eurasian snow cover and the Indian monsoon and not the whole Asian monsoon. The relationship between Eurasian winter snow cover and the summer rainfall in China is quite complicated as shown by Yang and Xu (1994). They divided China into five regions having similar characteristics of interannual variability in summer precipitation and found that, out of five regions, two are positively correlated and three are negatively correlated between Eurasian winter snow cover and summer precipitation.

In this paper the interannual variability of the south Asian and Indian summer monsoon rainfall, and their relationships with winter/spring snow mass over south Eurasia, are analysed based on a set of the present day climate simulations using the UK Universities' Global Atmospheric Modelling Programme (UGAMP) GCM. The snow mass is more physically meaningful than the snow-cover area for understanding the inverse snow–monsoon relationship because the former is a better measure of the total precipitation that finally results in soil moisture memory while the latter only reflects the change in surface albedo. Therefore, we will concentrate on the relationship between the snow mass and monsoon, although snow mass and snow-cover area in our model are strongly positively correlated simultaneously.

In section 2 the UGAMP GCM and the experiments analysed in this study are described briefly. The simulated mean monsoons are discussed in section 3. The snow mass and monsoon precipitation relationship is explored in section 4. Sensitivity of the result to the land surface parametrization schemes and the effect of varying SSTs are also discussed in this section. The composite analysis is performed in section 5 to highlight the physical mechanism responsible for the snow mass–monsoon relationship in the model. Sensitivity of the result to model horizontal resolutions is discussed in section 6, and the main results are summarized in section 7.

2. The model

The UGAMP GCM is based on the forecast model of the European Centre for Medium-Range Weather Forecasts (ECMWF). The version we are using is nearly identical to that currently being used in the Atmospheric Model Intercomparison Project (AMIP) which is described in detail by Slingo et al. (1994).

It is a spectral model with a hybrid sigma/pressure coordinate in the vertical, using a triangular truncation. The physical parametrizations in the model are described by Slingo
TABLE 1. VARIOUS EXPERIMENTS AND LENGTH OF SIMULATIONS

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Description</th>
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<tbody>
<tr>
<td>CLSST-A</td>
<td>21-year simulation at T42 using A&amp;M climatological SSTs with the no-flux boundary condition at the bottom of the soil model</td>
</tr>
<tr>
<td>AMIP-A</td>
<td>10-year simulation at T42 using AMIP varying SSTs with the no-flux boundary condition at the bottom of the soil model</td>
</tr>
<tr>
<td>CLSST-B</td>
<td>10-year simulation at T42 using climatological AMIP SSTs with the fixed climatology at the bottom of the soil model</td>
</tr>
<tr>
<td>AMIP-B</td>
<td>10-year simulation at T42 using AMIP varying SSTs with the fixed climatology at the bottom of the soil model</td>
</tr>
<tr>
<td>CLSST-T31A</td>
<td>10-year simulation at T31 using A&amp;M climatological SSTs with the no-flux boundary condition at the bottom of the soil model</td>
</tr>
<tr>
<td>CLSST-T21A</td>
<td>10-year simulation at T21 using A&amp;M climatological SSTs with the no-flux boundary condition at the bottom of the soil model</td>
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See text for explanation of the SSTs used.

et al. (1994). For the horizontal resolution at total wave number 42 (T42), the physical parametrizations are evaluated on a longitude/latitude grid of 128 by 64 points, where the mesh size is approximately 2.8°. For the horizontal resolution at T31 (T21), they are evaluated on a longitude/latitude grid of 96 (64) by 48 (32) points, where the mesh size is approximately 3.8° (5.6°). The model has 19 levels in the vertical, five of which are within the lowest 150 hPa of the atmosphere (i.e. in the boundary layer).

The land surface temperature and moisture content are calculated using a three-layer diffusive model. Either a no-flux boundary condition or fixed climatology at the bottom of the soil model is used. The no-flux boundary condition allows the surface temperature and soil moisture to respond fully to the forcing rather than being tied to some current climatology. Such a soil model has been used by Valdes and Hall (1994), Dong and Valdes (1995), and Hall et al. (1996) for paleoclimate studies.

An interactive surface hydrology is employed. Snow depth is computed as the balance of snow fall, snow melting and sublimation. Snow melting occurs when the surface energy balance requires the temperature to be raised above the freezing point. Surface albedo and roughness length are prescribed. When there is snow cover the surface albedo is weighted by snow albedo, which is snow-depth dependent. The deep snow (at infinite depth) albedo is 0.8. SST in the model is specified and over water the surface albedo is 0.07.

The sea-ice edge and sea-ice temperature are prescribed in the simulations with the fixed bottom-boundary conditions. However, they are treated differently in the simulations with the no-flux bottom-boundary condition, in which the sea-ice edge is prescribed as the −2 °C contour in the SST, and on sea ice itself the surface temperature is calculated using a simple sea-ice model. This is a purely thermodynamic model in which heat is stored by and diffused vertically across a slab of ice which is two metres thick. Over sea ice the surface albedo is 0.55.

A total of six simulations have been analysed. The experiment designs and lengths of simulations are summarized in Table 1. The Alexander and Mobley (1976) climatological SSTs are referred to as A&M climatological SSTs and the observed SSTs for the decade 1979–88 are referred to as AMIP varying SSTs.

3. THE MEAN CLIMATOLOGY

(a) Mean monsoon

The pattern and strength of the lower tropospheric south-west flow over the Arabian Sea and across the Indian peninsula, and upper tropospheric easterlies over South Asia, are
important features of the Asian summer monsoon. The time-mean lower and upper tropospheric flows for the various simulations at T42 are shown in Fig. 1 and Fig. 2 respectively, together with the corresponding ECMWF analysis. The model-simulated pattern of low-level south-westerlies over the Arabian Sea is similar to the analysis. However, the strength of the entire cross-equatorial circulation in the simulations is about 2 to 5 m s\(^{-1}\) weaker than the analysis, but with the CLSST-A and AMIP-A simulations giving a better agreement with the analysis. The easterly trade winds across the southern Indian Ocean and the anticyclonic circulation around the Arabian Sea are well represented in all simulations. The southerly jet over south-east China, which is related to the south-east Asian monsoon, is slightly stronger in the simulations than the observations, whereas the magnitude of the north-east and south-west trade winds over Africa and their convergence at 5°N–15°N are fairly well simulated. The main error in the CLSST-A simulation is the extension of the westerlies over the Philippines out into the western Pacific, which is associated with the development of a trough in this region. The analysis and the other three simulations show no penetration of westerlies in excess of 5 m s\(^{-1}\) east of the Philippines.

At upper levels, the southern Asian anticyclone associated with the monsoon is also in good agreement with the ECMWF analysis, its centre being located at 28°N over south Asia. The patterns of the easterlies over the Arabian Sea and Indian peninsula are similar to the analysis. However, the magnitude is underestimated by 2 to 5 m s\(^{-1}\) in the CLSST-B and AMIP-B simulations. The westerlies over the equatorial western Pacific are about 5 m s\(^{-1}\) stronger in the CLSST-A simulation. The errors in the June to August (JJA) circulation in the CLSST-A simulation are similar to several others GCM simulations, e.g. the National Center for Atmospheric Research Community Climate Model 1 (Meehl 1994) and the UK Meteorological Office GCM (Hewitt and Mitchell 1996). It seems that these errors may be due to the SSTs, rather than the land surface parameterization scheme in the UGAMP GCM, because in the AMIP-A simulation the circulation over the western Pacific is more consistent with the CLSST-B and AMIP-B simulations. The reason why the model produces such a large error in circulation over the equatorial western Pacific using the A&M climatological SSTs is unclear. It needs further investigation and is beyond the scope of this paper.

The time-averaged precipitation in various simulations over tropical and subtropical regions in the eastern hemisphere for the June to September (JAS) average is given in Fig. 3, together with the merged analysis of precipitation of Xie and Arkin (1996) and Legates and Willmott (1990) climatology. The standard deviations of the interannual variability in precipitation in JAS are given in Fig. 4.

Generally, the overall patterns of the precipitation in the four simulations at T42 are similar. The areas of major precipitation are simulated with both land surface schemes either using the climatological SSTs or AMIP varying SSTs. The precipitation rate over south-east Asia ranges from 4 mm day\(^{-1}\) to 10 mm day\(^{-1}\), which is in agreement with observation. However, the model produces a local precipitation maximum there, which is unrealistic. The rainfall over north-east India and over the west part of the Bay of Bengal is underestimated. The simulated precipitation rate is about 2 mm day\(^{-1}\), while observations give 4–8 mm day\(^{-1}\). The weaker simulation of Indian summer monsoon precipitation may be due to the use of envelope orography in the model. Fennessey et al. (1994) show that the replacement of the envelope orography with a mean orography in the Center for Ocean–Land Atmosphere Studies GCM results in a much more realistic simulation of Indian rainfall. Consistent with the simulated circulation, the model precipitation over north India, though weak, is stronger in the CLSST-A and AMIP-A simulations than in the CLSST-B and AMIP-B simulations, implying the importance of land surface schemes.
Figure 1. The mean 850 hPa wind distribution in June to August for (a) CLSST-A, (b) AMIP-A, (c) CLSST-B and (d) AMIP-B simulations (see Table 1), and (e) the ECMWF analysis. Light and dark shading indicate regions where winds speed are greater than 5 m s\(^{-1}\) and 10 m s\(^{-1}\) respectively.
Figure 2. As in Fig. 1, but for the wind at 200 hPa. Light and dark shading indicates regions where wind speeds are greater than 20 m s⁻¹ and 30 m s⁻¹ respectively.
Figure 3. The mean precipitation in June to September for (a) CLSST-A, (b) AMIP-A, (c) CLSST-B, and (d) AMIP-B simulations (see Table 1), (e) the merged precipitation of Xie and Arkin (1996), and (f) Legates and Willmott (1990) climatology. Contours are at 0.5, 1.0, 2.0, 4.0 and 8.0 mm day\(^{-1}\), then increasing by 4.0 mm day\(^{-1}\). The regions with precipitation rate greater than 4.0 mm day\(^{-1}\), 8.0 mm day\(^{-1}\), and 16.0 mm day\(^{-1}\) are shaded lightly, intermediately, and heavily.

Figure 4 indicates that the interannual variability in precipitation over central and north India in the CLSST-A and AMIP-A simulations (Fig. 4(a) and Fig. 4(b)) is close to that based on the data of Xie and Arkin (1996). The F-test indicates that the interannual variability is not different from that based on observation at the 2% significance level. The interannual variability of JJAS precipitation over Indian and south Asian areas in the AMIP-A simulation (Fig. 4(b)) is greater than that in the CLSST-A simulation (Fig. 4(a)), indicating a contribution to the interannual variability of precipitation over the two regions by the interannual variability of SSTs. The interannual variability in precipitation over central and north India is underestimated in the CLSST-B and AMIP-B simulations (Fig. 4(c) and Fig. 4(d)). It should also be noted that the amplitudes of the interannual
variability of precipitation over the Indian and south Asian areas in those two simulations are comparable. These imply a constraint on the variability in precipitation both through local and remote effects by fixing deep-soil temperature and moisture. Over the tip of the Indian peninsula and a large area over East Asia, the interannual variability in precipitation is overestimated in all simulations. This is also true over large areas of the ocean.

The seasonal variations of the mean precipitation in the simulations and observations over India, north India and south India (land points only) are displayed in Fig. 5. Generally, the seasonal variations of the area-averaged precipitation in the simulations over north India are in agreement with Xie and Arkin (1996). Legates and Willmott climatology gives peak monsoon precipitation about 2.0 mm day$^{-1}$ higher than Xie and Arkin data. The major errors in the seasonal precipitation variation are that precipitation in June and September
Figure 5. Seasonal variations of precipitation averaged over (a) India, (b) north India, and (c) south India for various simulations (see Table 1 for details) and the merged precipitation of Xie and Arkin (1996) and the Legates and Willmott (1990) climatology.
in the simulations is considerably smaller than the observation. These errors are larger in the CLSST-B and AMIP-B simulations.

The simulated seasonal variation in precipitation over south India is less consistent with observation, with the duration of the rain season about two months short and the monsoon withdrawal a few months early in the simulations. The maximum precipitation occurs in June rather than July. After June, precipitation in the simulations decreases significantly while observations indicate it is maintained until November.

(b) Snow cover and mass

Generally, the characteristics of the seasonal evolution of snow cover are simulated reliably although snow cover is slightly overestimated in transition seasons (not shown). However, the interannual variation of the snow-cover area is underestimated in all simulations (not shown). The snow mass is well simulated in winter and summer seasons, but not in the transition seasons in which it is overestimated in all simulations. It is a common phenomenon that there is large discrepancy between observed and simulated snow mass in most GCMs in the transition seasons (Foster et al. 1996). However, one must bear in mind that the model verification of snow mass has been hindered by a lack of high-quality observation data. Some of the discrepancy may be due to snow density, which is assumed to be 300 kg m\(^{-3}\) in order to get snow mass from snow depth in observations. The overestimation of snow mass in the CLSST-B and AMIP-B simulations is more serious in the winter and spring season. This may be partly responsible for the large-scale mean summer monsoon circulation differences highlighted above between the simulations with the no-flux boundary condition and those with the fixed bottom-boundary conditions.

Figure 6 shows the annual cycle of snow mass over south Eurasia and its standard deviation of interannual variability for the various simulations at T42. Generally, the main features are similar to those of the total Eurasian snow mass. The standard deviation of snow mass in the CLSST-A and AMIP-A simulations in spring and early summer is about 15% to 40% of the time-mean snow mass while it is only about 10% to 20% of its time-mean snow mass in the CLSST-B and AMIP-B simulations. Smaller interannual variability in snow mass in the pre-summer monsoon season in the CLSST-B and AMIP-B simulations is associated with smaller interannual variations in precipitation, implying that the prescribed deep-soil properties provide a strong constraint on the interannual variability of land surface hydrology which plays an important role in the variability of the monsoon circulation.

Figure 6 also suggests that the effect of the interannual variability of SSTs on the interannual variability of snow mass is smaller than the changes of interannual variability of snow mass due to a change in the land surface parametrization scheme. However, one must bear in mind that different SST data sets are used in the CLSST-A and AMIP-A simulations.

In summary, the model does a reasonable job at simulating the main features of the Asian monsoon. Good aspects of the simulation are:

- Low-level circulation over Africa and Asia.
- Upper-level Asian anticyclone.
- Precipitation over China and north India.
- Interannual variability in precipitation over north India.

Poor aspects of the simulation are:

- Precipitation over south India and the Bay of Bengal.
- Seasonal evolution of precipitation over south India.
4. ENSO, WINTER/SPRING SNOW MASS AND SUMMER RAINFALL RELATIONSHIPS

(a) Winter/spring snow mass and Indian summer monsoon precipitation

Table 2 gives the correlation coefficients between Indian rainfall in JJAS and JJ and snow mass over south Eurasia in February to May (FMAM) for all simulations. In calculating correlation coefficients, the first year is not included because in all simulations there is a significant increase in snow mass from year 1 to year 2, indicating that the imposed initial snow mass is different from the model’s own climatology. For 18 (7) degrees of freedom, the values of the correlation coefficients that are significant above zero at the 10% and 5% significance levels are $-0.38 (-0.58)$, and $-0.44 (-0.67)$, respectively. It can be seen that the correlations between snow mass and precipitation in early summer over India and north India are negative in the CLSST-A and AMIP-A simulations and that they are statistically significant. Generally, no apparent relationship occurs between the snow mass and rainfall over south India. The lack of correlation in south India may be a symptom of the relatively poor model simulation of south Indian rainfall. Similarly, the lack of statistically significant correlations in August and September (not shown) in north
India may be related to early withdrawal of the simulated monsoon. Table 2 also indicates that the snow–monsoon relationship is not affected by the interannual variation of SSTs in the UGAMP model. However, the snow mass–monsoon precipitation relationship is weak in the CLSST-B and AMIP-B simulations.

The above results indicate that the correlation between the winter to spring snow mass and summer precipitation over India in the model simulation is sensitive to the land surface parametrization scheme. The land parametrization scheme with the fixed bottom-boundary conditions, in which the deep-soil moisture and deep-soil temperature are imposed, provides a strong constraint on the interannual variations in land surface properties which may play an important active role in the monsoon variability.

(b) **ENSO, snow mass and Indian summer precipitation**

Relationships between the ENSO and Indian summer monsoon have been documented by many previous studies. These include Rasmusson and Carpenter (1983); Shukla and Paolino (1983); Ropelewski and Halpert (1987); Khandekar (1991); Webster and Yang (1992); Ju and Slingo (1995); and Yang (1996). In brief, most of these studies showed that deficient (excessive) Indian summer monsoon precipitation is associated with El Niño (La Niña) events. A recent study by Yang (1996), based on the observational data, indicates an ENSO, Eurasian winter snow cover, and Indian summer monsoon precipitation relationship. More (less) Eurasian winter snow cover occurs during El Niño (La Niña) winters.

The relationship between monsoon precipitation and SST is investigated in the AMIP-A and AMIP-B simulations. Teleconnection patterns of the correlation between the SSTs in JJA and the precipitation over India in JIAS in the two simulations and in that based on Xie and Arkin (1996) are shown in Fig. 7. It indicates that the relationship is negative over the east central Pacific and a large area over the Indian Ocean, and positive over the west Pacific. Observations give a similar pattern although the relationship is stronger in the model simulations. Figure 7 also indicates that the negative relationship over the central east Pacific is stronger in the AMIP-A simulation. The correlation coefficients greater than 0.58 or less than −0.58 in Fig. 7 are statistically significant at the 10% significance level.

In the AMIP-A simulation the relationship between the snow mass in FMAM and summer monsoon precipitation over India is quite high, as indicated in Table 2. This relationship is related to SST anomalies over the tropical west and east Pacific, which appear to affect the pre-monsoon circulation, inducing snowfall changes, which in turn have a feedback on the summer monsoon circulation. This is explained by Fig. 8, which shows

### Table 2. Correlation coefficients of the snow mass over South Eurasia (0°N–50°N, 60°E–120°E) in February to May with precipitation in the following summer (June to September) and early summer (June and July) over India (7°N–30°N, 70°E–90°E), North India (20°N–30°N, 70°E–90°E), and South India (7°N–20°N, 70°E–90°E) (land only) in the various simulations (see Table 1)

<table>
<thead>
<tr>
<th></th>
<th>June to September</th>
<th>June and July</th>
</tr>
</thead>
<tbody>
<tr>
<td>India</td>
<td>−0.465</td>
<td>−0.660</td>
</tr>
<tr>
<td>N. India</td>
<td>−0.511</td>
<td>−0.430</td>
</tr>
<tr>
<td>S. India</td>
<td>−0.055</td>
<td>−0.620</td>
</tr>
<tr>
<td>India</td>
<td>−0.516</td>
<td>−0.630</td>
</tr>
<tr>
<td>N. India</td>
<td>−0.055</td>
<td>−0.430</td>
</tr>
<tr>
<td>S. India</td>
<td>−0.603</td>
<td>−0.039</td>
</tr>
<tr>
<td>CLSST-A</td>
<td>−0.567</td>
<td>−0.690</td>
</tr>
<tr>
<td>AMIP-A</td>
<td>−0.247</td>
<td>−0.064</td>
</tr>
<tr>
<td>AMIP-B</td>
<td>−0.562</td>
<td>−0.650</td>
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<tr>
<td>CLSST-T31A</td>
<td>−0.552</td>
<td>−0.435</td>
</tr>
<tr>
<td>CLSST-T21A</td>
<td>0.466</td>
<td>0.206</td>
</tr>
</tbody>
</table>
the teleconnection pattern of the correlation coefficients between SSTs in the previous December to February (DJF) and snow mass over south Eurasia in FMAM. It clearly indicates a positive correlation over the tropical central and east Pacific and Indian Ocean and negative correlation over the tropical west Pacific. These correlations over large areas are statistically significant at the 10% significance level. These results suggest that premonsoon SST anomalies over the three regions clearly have an effect on the monsoon precipitation. Negative SST anomalies in the western Pacific and positive SST anomalies over the eastern and central Pacific in the previous winter and spring are favourable to heavy snowfall over south Eurasia in spring, which is associated with the weakened Indian summer rainfall, and visa versa. It is interesting that the snow mass over south Eurasia has been produced in such a way that the SST–monsoon relationship holds, i.e. there is a clear ENSO–snow–monsoon relationship. However, this is only true in the AMIP-A simulation. A detailed monsoon–ENSO relationship in the AMIP-B simulation was presented by Ju and Slingo (1995), which shows a systematic shift in the subtropical jet in the proceeding
winter associated with El Niño. The main purpose of this paper is an investigation of the snow–monsoon relationship in the various simulations with the UGAMP GCM. Detailed discussion of the circulation difference between strong monsoon and weak monsoon years in the previous winter and spring is beyond the scope of this paper.

(c) Spatial patterns of the snow–monsoon precipitation relationship

The distribution of monsoon precipitation over India and south Asia is highly variable in time and space. Therefore, the correlation coefficient between the snow mass over south Eurasia in FMAM and precipitation in JJAS is calculated over the whole globe. The results over part of the eastern northern hemisphere are shown in Fig. 9 for the CLSST-A and AMIP-A simulations. Generally, the overall pattern of the correlation is similar between the two simulations. There is a negative correlation in most regions over India except for the southern tip of the peninsula. The maximum negative correlation coefficients are found over north India, between 20°N and the foothills of the Himalayas. This spatial distribution is in good agreement with observational studies by Dey and Bhanu Kumar (1983). Once again, the correlation coefficients greater than 0.38 or less than −0.38 in Fig. 9(a) and greater than 0.58 or less than −0.58 in Fig. 9(b) are significant from zero at the 10% significance level. These relationships may provide useful information for seasonal and interannual prediction of the Indian summer monsoon precipitation, especially over north India and the foothills of the Himalayas where the correlation is the strongest.

5. Composite analysis

The relationships we discussed in the previous section are important for understanding the physical mechanism for modulating the monsoon. It involves the snow mass over
Eurasia, the subsequent snow melt and carry over of meltwater as soil moisture, the reduction of surface heating induced by evaporation, and a consequent weakening of the monsoon circulation in the heavy snow years. In this section, comparisons of the monsoon circulation between the heavy and light snow years in the CLSST-A simulation are presented by composite analysis of three heavy and four light snow years in which snow mass anomalies in FMAM exceed one standard deviation. One needs to bear in mind that in this simulation there is no interannual variation in SSTs. This section highlights the importance of land surface processes in the monsoon variability.

The seasonal variations of snowfall, snow mass, and surface soil moisture over south Eurasia are given in Fig. 10 for the heavy and light snow years, together with the corresponding standard deviations of interannual variability. The largest differences in the snowfall and snow mass between the heavy and light snow years occur in winter and spring, in which the differences are three times larger than the corresponding standard

Figure 9. Spatial distributions of correlation coefficients between snow mass in February to May over south Eurasia and precipitation in June to September for (a) CLSST-A, and (b) AMIP-A (see Table 1).
Figure 10. Seasonal variations of the composite for (a) snowfall, (b) snow mass, and (c) surface soil moisture over south Eurasia for the heavy and light snow years in the CLSST-A simulation (see Table 1) and the corresponding standard deviations of interannual variability.
Figure 11. Spatial distributions of differences in (a) snow depth and (b) soil moisture between the composite heavy and light snow years in February to May in the CLSST-A simulation (see Table 1). In (a) contours are for 1.0, 2.5, 5.0 and 10.0 mm in liquid water equivalence, then sequentially doubled; for (b) contours are for 0.5, 1.0 and 2.5 mm in liquid water equivalence. Negative values are shown dashed. Shading indicates differences are statistically significant at the 5% significance level using the two-sided Student's $t$-test.

deviations, and these differences are statistically significant at the 5% significance level. As suggested in Fig. 10, the rate at which snow is melted differs in the heavy and light snow years, especially during the period from late spring to early summer. Snow melt rate in the heavy snow years during this period is higher than that in the light snow years. It is important to note that differences in snow mass described here are smaller than the imposed snow differences by Yasunari et al. (1991), Vernekar et al. (1995) and Douville and Royer (1996).

Heavy snowfall and greater snow mass are also associated with greater soil moisture from winter to early spring. Only after May is the soil moisture over south Eurasia higher in the light snow years due to enhanced monsoon precipitation over north India and the foothills of the Himalayas from early summer.

Figure 11 gives the differences in snow depth and soil moisture in FMAM between the heavy and the light snow years. Before the summer season, there is more snow mass over Eurasia, especially over the Tibetan Plateau in the heavy snow years. Higher amounts of snow mass over south Eurasia in winter and early spring requires large amounts of energy to melt it during late spring and early summer which leads to high soil moisture as well. In fact, in our simulations, the snow mass and soil moisture over south Eurasia are highly positively correlated simultaneously in spring months.
The intensity of the Asian summer monsoon circulation to a large extent is proportional to the magnitude of the thermal contrast between the Eurasian continent and adjacent oceans. The higher albedo of snow reduces the absorption of incoming solar radiation, which in turn reduces the heating of the land surface of the continent from the spring to summer season. In addition, snow mass not only consumes the solar energy during snowmelt, but also reduces the heating of the surface after snowmelt by moistening the soil with melted snow. In heavy snow years, both the snow-albedo effect and snow-hydrological effect will be enhanced.

Due to both the snow-albedo effect and snow-hydrological effect, the surface temperature over large parts of Eurasia in spring is 1.0 to 2.0 degC colder in the heavy snow years. Temperature from the lower troposphere to mid-troposphere is also colder due to the reduction in sensible-heat fluxes. This is illustrated in Fig. 12. The magnitude of the meridional temperature gradient in late spring and early summer is of critical importance to the establishment of the summer Asian monsoon. The reduced meridional temperature gradient in spring in the heavy snowfall year delays and weakens the Asian summer monsoon circulation, resulting in a decrease in precipitation and latent-heat release due to condensation, which in turn has a positive feedback on the weaker monsoon circulation.

The outgoing long-wave radiation (OLR) anomalies from March to September between the heavy and light snow years are shown in Fig. 13. In general, positive OLR anomalies imply less cloudiness and/or lower cloud tops and, in regions where convection already exists, are assumed to be associated with suppressed convection. Negative anomalies are usually indicative of enhanced convection. The negative anomalies from March to April over south Asia may indicate that the surface in the heavy snow years is colder because at this time of the year there is no deep convection over there. After May, in June and July, the convection over south Asia and north India is significantly suppressed in the heavy snow years. In August, the OLR anomalies show a dipole structure with enhanced convection over central India and the Bay of Bengal and suppressed convection north of it. The convection over north India is enhanced in the heavy snow years in September, resulting in a slightly enhanced precipitation over north India as shown in Fig. 16(b).

The differences of the JJA mean monsoon circulation between the heavy and light snow years are shown in Fig. 14. The north–east anomalies over the Arabian Sea in the lower troposphere indicate that the south-west monsoon flows are weakened by 2.0 m s$^{-1}$ there during JJA in the heavy snow years relative to that in the light snow years. However, the westerlies from the Indian Ocean across south-east Asia into the west Pacific are stronger, and over the west Pacific low-level convergence is enhanced in the heavy snow years. The upper-level easterlies over the Arabian Sea and Indian subcontinent are also slightly weaker in the heavy snow years. The differences in both the upper and lower troposphere suggest a weaker Asian summer monsoon circulation is associated with the heavy snow years.

The velocity potential anomalies in the upper troposphere suggest that the divergence is weakened over the Arabian Sea and enhanced over the west Pacific in the heavy snow years. It is interesting to note that the pattern is the opposite of that of the velocity potential difference between 1988 (La Niña) and 1987 (El Niño) displayed by Palmer et al. (1992), but its magnitude is only half. Detailed circulation contrasts between El Niño and La Niña years are also discussed by Ju and Slingo (1995) based on both the model and observations.

The OLR anomalies between the composite heavy and light snow years indicate that the convective activity in JJA over the Arabian Sea and north India are weaker during the heavy snow years.

The spatial pattern of the precipitation anomalies in JJAS between the heavy and light snow years is shown in Fig. 15. It indicates that the summer precipitation over India is
Figure 12. Spatial distributions of differences in temperature in March to May between the composite heavy and light snow years in the CLSST-A simulation (see Table 1) for (a) surface temperature, (b) temperature at 850 hPa, and (c) temperature at 500 hPa. Contours are at 0.5, 1.0, 2.0 degC, then increasing by 1.0 degC with negative anomalies dashed. Shading indicates that differences are statistically significant at the 5% significance level using the two-sided Student's t-test.
Figure 13. The outgoing long-wave radiation anomalies between the heavy and light snow years from March to September in the CLSST-A simulation (see Table 1). Contour intervals are 5.0 W m⁻² with negative values dashed. Shading indicates the differences are statistically significant at the 5% significance level using the two-sided Student's t-test.
Figure 14. Spatial distributions of differences in (a) 850 hPa wind, (b) 200 hPa wind, (c) 200 hPa velocity potential, and (d) outgoing long-wave radiation between the composite heavy and light snow years in the CLSST-A simulation (see Table 1). Contour intervals are $5 \times 10^{8}$ m$^2$s$^{-1}$ in (c) and 5.0 W m$^{-2}$ in (d) with negative values dashed. Shading indicates the differences are statistically significant at the 5% significance level using the two-sided Student's t-test.
0.5 mm day\(^{-1}\) to 3.0 mm day\(^{-1}\) less in the heavy snow years. This corresponds to a decrease of about 10% to 30% relative to that in the light snow years. Precipitation over north-east China and central China is also less in the heavy snow years. However, precipitation over east China and north China is enhanced. The positive anomalies north of the Tibetan Plateau are probably associated with the positive snow-hydrological effect. Heavy snow in winter and early spring moistens the soil, enhancing evaporation and increasing water vapour in summer, which in turn induces an increase in precipitation. The precipitation anomalies over China are consistent with the observation study by Yang and Xu (1994), in which it was suggested that the excess Eurasian winter snow cover is associated with less precipitation over north-east and central China and more precipitation over south-east and north China in the summer.

The seasonal variations of precipitation over India, north India, and south India are shown in Fig. 16. They clearly indicate the increased precipitation in early summer in the light snow years relative to that in the heavy snow years. They also suggest that the onset of the monsoonal precipitation over north India is earlier in the light snow years. It is interesting to note that the largest difference in precipitation occurs during the onset of the monsoon and early summer but does not persist through the whole season. By August, precipitation over India is very similar for the heavy and light snow years, and precipitation in the retreat phase is almost identical. Ju and Slingo (1995) also noted that the influence of the ENSO on monsoon circulation is not seen in the latter part of the annual cycle in either the model or the analysis results.

6. Sensitivity to horizontal resolution

The simulated mean JJAS precipitation and its standard deviations of the interannual variability at T31 and T21 horizontal resolutions are shown in Fig. 17. In comparison with observation and simulations at T42 (Figs. 3 and 4), it is seen that the precipitation pattern in the model at T21 is vastly different from those simulated at higher resolution over India and adjacent regions. The sharp decrease in precipitation to the east of India, the rainshadow effect in the lee of the mountains, simulated at T42 and seen in observations, is totally missed in the CLSST-T21A simulation, consistent with the study of Sperber et al. (1994) and Royer et al. (1994). There is a significant pattern change in precipitation distribution over India and its adjacent regions from T21 to T31, with the CLSST-T21A
Figure 16. Seasonal variations of precipitation averaged over (a) India, (b) north India, and (c) south India for the composite heavy and light snow years, and the corresponding standard deviations of its interannual variation in the CLSST-A simulation (see Table 1).
Figure 17. (a) and (b). The mean precipitation in June to September and (c) and (d) the corresponding standard deviations of its interannual variability in the simulations at T31 and T21. (a) and (c) the CLSST-T31A and (b) and (d) the CLSST-T21A simulations (see Table 1). In (a) and (b) the contours are at 0.5, 1.0, 2.0, 4.0 and 8.0 mm day$^{-1}$, then increasing by 4.0 mm day$^{-1}$. The regions with precipitation rate greater than 4.0 mm day$^{-1}$, 8.0 mm day$^{-1}$, and 16.0 mm day$^{-1}$ are shaded lightly, intermediately, and heavily. In (c) and (d) the contours are at 0.5, 1.0 and 2.0 mm day$^{-1}$, then increasing by 2.00 mm day$^{-1}$. The regions with standard deviation greater than 2.0 mm day$^{-1}$, 4.0 mm day$^{-1}$, and 6.0 mm day$^{-1}$ are shaded lightly, intermediately and heavily.

simulation giving a precipitation maximum over south-east India which is unrealistic, while the CLSST-T31A simulation is more or less similar to those in the CLSST-A simulation. Seasonal evolution of snow mass over south Eurasia shown in Fig. 18 indicates that winter/spring snow melts completely one month earlier in the CLSST-T31A simulation than in the CLSST-A simulation. This may be partly responsible for the fact that the maxima in the mean precipitation in July and August over north India in the CLSST-T31A simulation is 0.5–1.5 mm day$^{-1}$ greater than that in the CLSST-A simulation. The interannual variability in JIAS precipitation in the CLSST-T31A simulation is more or less similar to that in the CLSST-A simulation. However, the CLSST-T21A simulation indicates the largest interannual variability over central India, which is not present in either the simulations at higher resolution or in observations.

The inverse relationship between snow mass in winter/spring over south Eurasia and summer Indian monsoon precipitation in the CLSST-T31A simulation is illustrated in Table 2. The characteristics are similar to those in the CLSST-A simulation. The spatial pattern of the correlation coefficient (not shown) suggests that a strong negative correlation occurs over north India and the foothills of the Himalayas, which is also in agreement with that in the CLSST-A simulation.

However, in the CLSST-T21A simulation, there is no such negative relationship between the snow mass in winter/spring and summer Indian precipitation, as suggested by Table 2. Instead, it shows a generally positive correlation between the two quantities. This
Figure 18. (a) Seasonal variations of snow mass over South Eurasia and (b) the corresponding standard deviations of its interannual variability in the CLSST-A, CLSST-T31A, and CLSST-T21A simulations (see Table 1).

is in contradiction with the results in the simulations at higher horizontal resolution. The cause of the change in sign of the snow mass–monsoon precipitation correlation from the simulation at T31 to the simulation at T21 is unclear. It may be related to the snow mass seasonal cycle and its interannual variability over South Eurasia. The snowmelt in spring in the CLSST-T21A simulation is generally much quicker than in the simulations at higher resolutions. The accumulated snow in the previous winter and spring melts completely by June and there is rather small interannual variability in snow mass from May to July in the CLSST-T21A simulation, which is in contrast with the simulations at higher resolution. Thus the simulated mean monsoon precipitation pattern, the seasonal cycle of snow mass, and its interannual variability in the CLSST-T21A are vastly different from those in the simulations at higher horizontal resolution. It is not expected that the interannual variability of monsoon precipitation can be correctly simulated, as Shukla and Fennessey (1994) suggested that a prerequisite to simulate the interannual variability of the Indian monsoon is that the mean monsoon is correctly simulated.
7. DISCUSSION AND SUMMARY

In this study the interannual variability of the Asian and Indian summer monsoon has been analysed based on various climate simulations using the UGAMP GCM. The results indicate that there is, generally, a negative correlation between the winter/spring south Eurasian snow mass and the subsequent summer monsoon precipitation over India, especially over north India. This relationship is sensitive to the land surface parametrization scheme, with the simulations using the no-flux bottom-boundary conditions giving a stronger negative correlation.

The result also hints at a link between SST anomalies in the previous winter/spring, and snow mass in the spring, and monsoon summer precipitation over India. The simulation with the no-flux boundary condition using the AMIP varying SSTs indicates an ENSO–snow mass–monsoon relationship. Snow mass is produced in such a way that the ENSO–monsoon relationship is maintained.

The impact of SST anomalies on the Asian summer monsoon may be both direct and indirect. The direct impact can be manifested by the simultaneous SST–monsoon relationship. The indirect impact of SST on the monsoon may be understood by its possible effect on the land surface processes in the Asian continent, which in turn cause variability in the following summer monsoon. Thus, the effects of SST anomalies and land surface processes are not necessarily mutually exclusive but rather work interactively to produce variations in the Asian summer monsoon. Results in this paper also indicate that warm SST anomalies over the central and east Pacific lead to increased snow mass and soil moisture in the winter and spring over south Eurasia, which has a significant positive feedback on the summer monsoon. It is hard to separate how much the influences are from SST direct effect and how much are from land surface processes induced by the SST anomalies. However, the results in the CLSST-A simulation (in which there is no interannual variations in SSTs) indicate that snow mass may have a significant impact on the summer monsoon circulation and precipitation over India.

The greater winter snowfall delays and weakens the heating of the Tibetan Plateau and overlying atmosphere, due to snow-albedo and snow-hydrological effects. As a result, the summer monsoon circulation is weaker in the heavy snow years than in the light snow years, at least in the early part of the season. The weaker monsoon circulation is associated with the reduced south-westerly summer monsoon winds over the Arabian Sea in the lower troposphere, decreased divergence over the Indian subcontinent in the upper troposphere, decreased convective activity over the Arabian Sea and Indian subcontinent, and reduced precipitation over India. Thus, the model results support the hypothesized snow–monsoon relationship and agree with observations.

Sensitivity studies to horizontal resolution suggest that the model at T21 resolution is unable to simulate the mean monsoon precipitation over India and the adjacent regions. This feature is similar to that of Sperber et al. (1994) using the ECMWF model in which it is concluded that on time-scales of a month or longer the most demonstrable differences in monsoon circulation occur between the simulation at T21 and that at T42. This study suggests that at least a T31 horizontal resolution is needed to simulate the interannual variability of the Indian summer monsoon, especially for the UGAMP model.

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