Asian monsoons in the future

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17.1 INTRODUCTION: GLOBAL ASPECTS

Climate models (specifically, global coupled atmosphere–ocean general circulation models (GCMs)) are the most useful tools for projecting future climate changes and variations across time. Several scenario experiments for future climate projections have been done recently, in addition to the 1% per year CO₂ increase experiments commonly used by the modeling community. These recent simulations used the marker scenarios identified by the Intergovernmental Panel on Climate Change (IPCC) Special Report on Emission Scenarios (SRES; IPCC, 2000) and shared many common results, including an increase in global mean surface temperature, an increase in global mean precipitation, greater increase in surface temperatures over land than over ocean, more temperature increases at high latitudes than at low latitudes, and more temperature increases in the northern hemisphere than in the southern hemisphere. However, there are still discrepancies among the models in terms of the geographical distribution of precipitation changes.

In order to get an idea of the global-scale temperature and precipitation trend and its uncertainty, it may be useful to use the 1% per year CO₂ increase experiments of the second phase of the Coupled Model Intercomparison Project (CMIP2; Meehl et al., 2000). Figures 17.1 and 17.2 show the global distribution trends of the annual mean surface air temperature and annual mean precipitation for the next 100 years, as calculated by 11 climate models using the 1% per year CO₂ increase scenario (Noda et al., 2003). The two figures show the mean values for the 11 models (the model ensemble mean) and the standard deviations among the models in order to illustrate the uncertainty.

The ensemble mean temperature trend clearly illustrates the differences in surface temperature changes already mentioned. Specifically, the regions with warming of at least 5°C per 100 years are seen only in the northern high latitudes, and the maximum temperature rise is seen in the Arctic region where sea ice
reduction is eminent, according to the future climate projections. Over the Eurasian and North American continents, the temperature rise in the east is larger than in the west. It is thought that the temperature increase is larger over these regions due to a strong snow/ice–albedo feedback effect, because the snow line is located at relatively low latitudes (Noda et al., 1996). The scatter among models was greater at higher latitudes than at lower latitudes, and in the southern hemisphere, the scatter was more pronounced around Antarctica.

The geographical distribution of predicted changes in annual mean surface temperatures by the end of the 21st century, as shown by the SRES-A2 and B2 scenario multimodel projections (IPCC, 2000), is quite similar to Figure 17.1. In these projections, the A2 and B2 storylines and scenarios reflect a very heterogeneous
world of self-reliance and preservation of local identities, and a world in which emphasis is on local solutions for economic, social, and environmental sustainability, respectively (IPCC, 2000). The average global mean surface air temperature increase is predicted to be 3.6°C over the next 100 years, but the increase over Asia is larger than the global average, and the increase over Japan in particular is 4–5°C (Noda et al., 2003). Zhao et al. (2004) assembled the surface temperature projections averaged across east Asia (15°–60°E and 70°–150°E), using seven GCMs with various scenarios, and predicted that the surface temperature will increase between 3.4°C and 5.5°C by 2099.

The global annual mean precipitation is also increasing (Allen and Ingram, 2002), and large precipitation increases can be seen in the tropical Pacific (Figure 17.2). Trends of increasing precipitation in the equatorial Pacific and Indian Ocean and of decreasing precipitation in the subtropical South Pacific and South Indian Ocean are evident, but the scatter among models for these regions is
large. Over the Eurasian continent, the models predicted increased precipitation, with a larger increase to the south-east of the Tibetan Plateau. Over the Mediterranean region, precipitation was predicted to decrease.

The intermodel standard deviation of the projected precipitation trend is very large over the tropical oceans. The scatter is also large over India, which suggests caution in accepting the quantitative accuracy of Indian monsoon projections. It is evident that consistency among models is lower for precipitation than for temperature. Over the south Asia and east Asia regions the IPCC (2001) compared model consistency in the seasonal mean precipitation change, and found that models with the SRES-A2 scenario were consistent in showing increased precipitation in the summer months (June, July, and August), but were inconsistent in the winter months (December, January, and February).

One example of an MRI-CGCM2 simulation (Figure 17.3) showed changes in surface air temperature and precipitation for both the summer and winter months (Noda et al., 2001; Ashrit et al., 2005). An ensemble mean of the three-member SRES-A2 scenario simulations at the end of the 21st century (2071–2100) is compared with present-day simulations. The summertime temperature change is predicted to be large over dry regions in the Middle East and western China but relatively small over the Asian monsoon area. There is a minimal temperature increase over north-western India, where precipitation and soil moisture both increase. Precipitation increases over India, the Bay of Bengal, and south China through Japan, while it decreases over northern China. The increase in precipitation change over the arid area of the Persian Gulf/Pakistan and north-western India is in contrast to the decrease around the Caspian Sea to the north. In the winter, the simulated surface temperature change is large at high latitudes and over the Tibetan Plateau. There is also a large temperature increase over the Okhotsk Sea due to reduced sea ice cover. An increase in precipitation is seen over the Maritime Continent and is associated with intensified easterly trade winds from the western Pacific to the Indian Ocean.

17.2 SUMMER MONSOON

17.2.1 South Asian monsoon

Many climate model studies reported possible changes in the south Asian summer monsoon due to CO₂ increase. Although a few studies reported that the south Asian summer monsoon becomes weak or that there is no significant precipitation change (Zhao and Kellogg, 1988; Lal et al., 1994, 1995; Lal and Singh, 2001), most models showed increases in both the seasonal mean precipitation and the interannual variability (Meehl and Washington, 1993; Bhaskaran et al., 1995; Kitoh et al., 1997; Giorgi and Francisco, 2000; Hu et al., 2000a,b; Douville et al., 2000a; May, 2002, 2004; Ashrit et al., 2003; Meehl and Arblaster, 2003).

One of the earlier results (Figure 17.4) plotted the south Asian (5°–40°N, 60°–100°E) summer precipitation for 1×CO₂ and 2×CO₂ equilibrium experiments
Figure 17.3. (a) JJA mean surface air temperature change between 2071–2100 and the present-day simulation (1971–2000) by the MRI-CGCM2. A three-member ensemble mean is shown with the SRES-A2 scenario. (b) JJA precipitation change. (c) JJA precipitation change as a ratio to the present value. (d–f) As in (a–c) except for DJF.

using the NCAR model (Meehl and Washington, 1993). Mean summertime precipitation increased 6.3% from 6.40 mm day$^{-1}$ to 6.80 mm day$^{-1}$, while standard deviation (interannual variability) increased from 0.49 mm day$^{-1}$ to 0.65 mm day$^{-1}$. Shown by the increase in both the mean value and standard deviation, the precipitation in wet years increased 11%; in dry years, the percentage changed slightly due to global warming. The researchers concluded that the precipitation increase is due to a larger increase of the Eurasian continental temperature than of the Indian Ocean temperature.
Using the 1% per year transient CO$_2$ increase experiments of the UKMO coupled model, Bhaskaran et al. (1995) found a 20% increase in the Indian summer monsoon precipitation at the time of CO$_2$ doubling. They concluded that the increase in precipitation is due to increased atmospheric moisture and increased land–sea temperature contrast. Also, from the 850 hPa wind system change, they found that monsoon circulation shifted to the north 10 degrees and was strengthened by 10%.

Kitoh et al. (1997) raised an issue about an apparent paradox between the south Asian summer monsoon’s increasing precipitation and its decreasing circulation intensity. Figure 17.5 shows the two monsoon indices at the time of CO$_2$ doubling and quadrupling according to the MRI-CGCM 1 (Kitoh et al., 1997): one is the all-India rainfall; the other is an averaged vertical zonal wind shear, which shows the difference in zonal wind between 850 hPa and 200 hPa, averaged over the region 40$^\circ$–110$^\circ$E, 5$^\circ$–20$^\circ$N (Webster and Yang, 1992). It is clear that the Indian summer monsoon precipitation increases with CO$_2$ increase. On the other hand, the monsoon wind index does not show a strengthening due to warming (Figure 17.5(b)); rather, the 850-hPa monsoon westerly jet and the 200-hPa easterly jet weakened in the summer months of July and August. Therefore, westerly wind shear becomes smaller with warming, which seems to be contradictory to the increase in the India summer monsoon precipitation. This paradox can be explained by the northward shift of the monsoon circulation. The lower tropospheric monsoon wind system shifts slightly to the north, thus leading to a weakening of the index defined over the 5$^\circ$–20$^\circ$N domain. The same is true for the CNRM GCM (Ashrit et al., 2003) and the MRI-CGCM2 (Ashrit et al., 2005).
Figure 17.5. (a) Seasonal cycle of precipitation averaged over land for 60°–100°E, 10°–30°N. The solid line is the 20-year average in the control experiment, the dashed (dotted) line represents the times of the 2 × CO₂ (4 × CO₂) in the MRI-CGCM1. (b) As in (a) except for zonal winds at 850 hPa and 200 hPa averaged for 40°–110°E, 5°–20°N. After Kitoh et al. (1997).

Another analysis (Figure 17.6) showed the total atmospheric moisture flux and its change in the MRI-CGCM2. Although the monsoon westerly wind from the Arabian Sea to India became weak, the total moisture flux increased due to the increase of the atmospheric moisture content resulting from temperature increase, thus leading to increased precipitation in India. Bhaskaran et al. (1995), Douville et al. (2000b), Ashrit et al. (2003), and Meehl and Arblaster (2003) also reached the same conclusion. May (2002, 2004) used a higher horizontal resolution model (with a T106 time-slice experiment with ECHAM4 AGCM using the 1970–1999 and 2060–
2099 sea surface temperature (SST) obtained by the T42 resolution ECHAM4/OPYC coupled GCM with IS92a scenario) and also showed a weakening of the monsoon circulation and an increase in the Indian summer monsoon precipitation due to strengthened moisture transport.

Using National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis data, Stephenson et al. (2001) showed a decrease in the monsoon westerly wind shear index (Webster and Yang, 1992) and the meridional wind shear index (Goswami et al., 1999) from −0.1 to −0.3% per year from 1958 to 1998, although the all-India summer precipitation showed no trend. Geological evidence based on the fossil record from the Arabian Sea shows an increase in the monsoon intensity over the past four centuries (Anderson et al., 2002), but this does not contradict the northward shift of monsoon circulations (Figure 17.6).

Using the Centre National de Recherches Meteorologiques (CNRM) model results, Ashrit et al. (2003) showed that the rate of increase in precipitable water is larger than the rate of increase in precipitation, suggesting a decrease in the moisture recycling ratio. Douville et al. (2002) conducted a detailed analysis of this rain efficiency decrease. Moreover, since rain efficiency depends on soil moisture, the
Figure 17.7. (a) Time series of the June–September precipitation anomalies in south Asia (40°–110°E, 0°–20°N) and its 11-year running mean of the IS92a run by the ECAHM4/OPYC3 CGCM. (b) Sliding variance of (a) using an 11-year window in the IS92a run (solid line) and control run (dashed line). (c, d) As in (a, b) except for the Niño 3 (150°–90°W, 5°S–5°N) SST.
Reproduced by permission of American Geophysical Union from figure 2 of Hu et al. (2000b).

Regional change in monsoon precipitation resulting from a CO2 increase is affected by the surface hydrological scheme (Douville et al., 2000b).

Most model results projected an increase in interannual variability associated with an increase in mean precipitation (e.g., Meehl and Washington, 1993; Bhaskaran et al., 1995; Kitoh et al., 1997; Hu et al., 2000b; Meehl and Arblaster, 2003). Bhaskaran et al. (1995) performed a frequency distribution analysis of daily precipitation in the UKMO model, and found an increase in the frequency of heavy rain. The result of the IS92a run by the MPI coupled model is shown in Figure 17.7 (Hu et al., 2000b). The researchers calculated the time series of the variance in an 11-year running window (Figure 17.7(b)), which showed an increase in interannual variability of the Indian summer monsoon rainfall after 2030. Hu et al. (2000b) related this to the increased variability of the tropical Pacific SST (El Niño variability; Figure 17.7(d)). Meehl and Arblaster (2003) explored the cause of change in monsoon precipitation variability by separately examining SST anomalies in the Pacific and in the Indian Ocean. They found that the increased variability in evaporation and precipitation in the Pacific due to increased SST influenced the South Asian monsoon variability through Walker circulation; the role of the Pacific Ocean is dominant and that of the Indian Ocean is secondary.
The effect of aerosols on the monsoon system is a major controversy. If the direct effect of aerosol increase is considered, then the increase in surface temperature over Asia is lessened because the aerosol particles reflect solar insolation. For this reason, the land–sea temperature difference becomes smaller than without the aerosol effect, and the south Asian summer monsoon becomes weaker (Meehl et al., 1996a). Using the UKMO model, Mitchell et al. (1995) and Mitchell and Johns (1997) reported that the south Asian (5°–30°N, 70°–105°E) summer monsoon rainfall decreased 5% when the direct effect of sulphate aerosol was considered, and it increased 5% when only the greenhouse gases (GHGs) were included. Rupak Kumar and Ashrit (2001) compared this UKMO model result with those of the MPI model. Although both models included the direct effect of sulphate aerosol, the MPI model showed an increase in precipitation while the UKMO model showed a decrease. Spatial patterns of monsoon precipitation were not sensitive to the magnitude of the forcing, if monsoon precipitation amount changes with warming. However, these model experiments have assumed that surface albedo change acts as a surrogate for the aerosol effect.

Other subsequent model results showed that the sulphate aerosol effect reduces the magnitude of precipitation change compared with the models with only GHGs, but the increase in south Asian summer monsoon precipitation remains (Roeckner et al., 1999; Emori et al., 1999). Furthermore, it was suggested that black carbon and dust, which absorb solar radiation, have a large influence on precipitation (Ramanathan et al., 2001). These aerosol particles float in the atmosphere in south Asia mostly in the dry season (November–May), but when the rainy season begins they are washed away. Since aerosols with high absorptivity (such as black carbon) absorb solar radiation in the lower atmosphere, the solar radiation that reaches the surface decreases. This causes precipitation to increase in winter over the Indian Ocean, and to decrease in the surrounding Indonesian region and the western Pacific Ocean (Chung et al., 2002). Black carbon also influences the summer monsoon, both in south Asia and east Asia (Menon et al., 2002).

17.2.2 East Asian monsoon

A number of studies also looked at future climate projections in east Asia (Zhao et al., 2004; and others discussed in this section). Using seven GCM results for 1 × CO₂ and 2 × CO₂ equilibrium experiments, Hulme et al. (1994) estimated that precipitation will increase over most of the east Asia region during all seasons. Giorgi and Francisco (2000) compared the local climatic change (2071–2100 compared with 1961–1990) using the IS92a transient experiment with five global climate models (HadCM2, CSIRO Mk2, CCCma/CGCM1, CCSR/NIES, and ECHAM/OPYC). They divided the World into 23 regions and investigated the temperature and precipitation changes in summer and winter. Although there were differences due to the aerosol effect and also among ensemble members, it was found that the intermodel difference was dominant. There was no summertime
precipitation change in east Asia when only the GHGs were considered, and no consensus among models was found when the effect of aerosol was included. Giorgi et al. (2001) summarized the nine global climate model results (CCCma/CGCM2, CSIRO Mk2, NCAR/CSM 1.3, ECHAM/OPYC, GFDL R30c, HadCM3, MRI-CGCM2, CCSR/NIES, and DOE PCM), which all used the SRES A2 and B2 scenarios and compared the periods of 2071–2100 and 1961–1990. According to this study, there was consensus among the models about the summer precipitation increase over east Asia (+5% to +20%). Hu et al. (2003) used the CMIP2 experiments with 16 models to show a trend of JJA increasing precipitation over almost all of China, except for the western part of Inner Mongolia at the time of CO₂ doubling. However, intermodel scatter was again very large.

There were some attempts to downscale climate change in east Asia by using regional climate models. Kato et al. (2001) investigated the east Asia climatic change in January and June at the time of CO₂ doubling using NCAR RegCM2.5 (50 km horizontal resolution), which is nested to NCAR CSM; one integration period covered ten years. They noted that in June, the subtropical anticyclone intensifies around 25°–30°N and extends westward; precipitation decreases near the Philippines, where the subtropical anticyclone belt intensifies, and increases around the Baiu rain area over south China and to the south of Japan. They also analyzed daily precipitation characteristics around Japan and showed that total precipitation increases, while there is no change in the frequency of wet days (daily precipitation greater than 1 mm day⁻¹), suggesting an increase in precipitation intensity. Gao et al. (2002) also showed an increase of heavy rain days in southern China at the time of CO₂ doubling by RegCM/China nested to CSIRO GCM.

Change in the above-mentioned Pacific anticyclone (i.e., the tendency for sea level pressure to become high in the region south of 30°N and become low in a 40°N–50°N belt, and the strengthening and southward movement of the Pacific anticyclone) has also appeared in the CO₂ increase experiment results by the MRI-CGCM1 (Kitoh et al., 1997). Nishimori and Kitoh (2002) developed a statistical downscaling method to project high spatial resolution regional precipitation changes based on large-scale circulation features obtained by GCMs, and applied it to MRI-CGCM1, CCSR/NIES, HadCM3, and ECHAM4/OPYC3. Although the GCM results are scattered among the models, their statistical downscaling method found common features such as an increase in precipitation over the northern part of Japan and a decrease over the Pacific coastal side of Japan, which is associated with the intensification and south-west movement of the Pacific subtropical anticyclone.

Both the regional climate models and the statistical downscaling method are influenced greatly by the accuracy of the GCM in reproducing the present-day climate (mean and interannual variability). In a control experiment using the present climate model, the reproducibility of actual summertime precipitation is poor, particularly over the western Pacific region; this is common with many GCMs (Kang et al., 2002). In order to predict the precipitation change in east Asia with sufficient accuracy, an improvement in the model climate is necessary.
17.2.3 Monsoon onset

Changes in the monsoon are not restricted to the seasonal mean precipitation but also include seasonality, such as monsoon onset and withdrawal dates. Figure 17.8(a,b) (color section) shows the distribution of onset dates of the summer rainy season according to the CMAP precipitation climatology and the present-day simulation by MRI-CGCM2 (Rajendran et al., 2004). Following Wang and LinHo (2002), onset is defined as the first pentad in which the five-day mean precipitation increases more than the January average precipitation at the grid point of more than 4 mm day$^{-1}$. Wang and LinHo (2002) observed that onset dates propagate over time from south-west to north-east over the Arabian Sea and from south-east to north-west over the Bay of Bengal. The MRI-CGCM2 effectively reproduced the observed characteristics of monsoon onset both in its spatial propagation and in its timing. The earliest onset takes place over the region around the Indo-China peninsula and the Bay of Bengal. The next onset occurs in mid-May over the South China Sea, and after that the onset dates propagate northward with the movement of the Meiyu/Baiu rain. Over the Philippine Sea and the western North Pacific, there are two jumps in the monsoon onset (Ueda and Yasunari, 1996). Although the movement of onset dates is generally well reproduced in the model, particularly over the Indian monsoon region, there remain some discrepancies in the western Pacific, showing that the models need further improvement in their climatology.

Figure 17.8(c) (color section) shows the predicted onset dates of the Asian summer rainy season at the end of the 21st century using the SRES-A2 scenario experiment of the MRI-CGCM2 (Kitoh, 2003). Earlier onset is more clearly seen in India than in other regions. Kitoh (2003) investigated regional differences in rainy seasons. In India, the model showed an overall increase in summertime precipitation, but the increase is larger in May and June, contributing to the earlier onset of the rainy season. In Indochina, although onset occurs one pentad earlier, the increase in the total amount of precipitation is not remarkable. On the other hand, the monsoon onset occurs later in the western North Pacific. In this area, the contrast between the decrease in May–June and the increase in July–August is remarkable, showing a clear onset. Around Japan, a change in the timing of onset is not seen, although precipitation increases. Although the Indian summer monsoon onset date becomes earlier in the MRI model, Bhaskaran and Mitchell (1998) presented another example showing a late Indian monsoon onset by the HadCM2 model where the seasonal average precipitation decreased. They also showed a delay in the onset in the western part of south-east Asia. Careful examination of other model results and the utilization of other methods is required.

In their projection of changes in the withdrawal of the rainy season, Uchiyama and Kitoh (2004) reported that the Baiu onset date does not change much, but the withdrawal near Japan clearly becomes later by the end of the 21st century in the MRI-CGCM2 SRES-A2 scenario experiment. They suggested that this delay of the Baiu withdrawal is caused by the large-scale circulation response associated with the El Niño-like mean SST change.
17.2.4 ENSO–monsoon relationship

The El Niño/Southern Oscillation (ENSO) affects interannual variability in the entire tropics through Walker circulation changes. Therefore, many are interested in how global warming changes the behavior of ENSO and the relationship between ENSO and the monsoon system. Many models so far have projected that the ‘time-average’ pattern of the tropical Pacific SST change due to global warming becomes El Niño-like (IPCC, 2001). That is, the SST rises more in the eastern tropical Pacific than in the western tropical Pacific, and the tropical Pacific rainfall distribution undergoes an eastward shift (Figure 17.2(a)).

So the question then is, what is the projected change in the amplitude, frequency, and spatial pattern of El Niño itself? IPCC (2001) concluded that the difference among the models is very large because in some models, ENSO activity becomes small and in others, it becomes large. This uncertainty is partly because these models are not able to reproduce ENSO with sufficient accuracy in their control experiment (Latif et al., 2001).

It has been known that there is a significant correlation between ENSO and the monsoon, shown by the analysis of observational data (below-normal Indian monsoon rainfall in El Niño years, and above-normal rainfall in La Niña years). Recent analysis revealed that this correlation has a remarkable decadal fluctuation (Krishna Kumar et al., 1999a). Variability in the relationship between the east Asian summer monsoon and ENSO is also reported (Wang, 2002). Moreover, since the correlation between ENSO and the Indian summer monsoon has recently collapsed, many hypotheses have looked for the reason; suggestions include decadal variability (Kripalani and Kulkarni, 1997b), a change in seasonality of the ENSO cycle (Kawamura et al., 2003), the Indian Ocean Dipole mode (Ashok et al., 2001), Atlantic Oscillation (Chang et al., 2001a), and global warming. With respect to global warming, one hypothesis is that the Walker circulation accompanying ENSO shifted south-eastward, reducing the downward motion that originally suppressed precipitation in the Indian region at the time of El Niño, resulting in a normal precipitation (Krishna Kumar et al., 1999b). Another explanation is that global warming raises the ground temperature of the Eurasian continent in the winter–spring season so that the temperature difference between the continent and the ocean becomes large; this causes more precipitation, and the Indian monsoon remains normal in spite of the occurrence of El Niño (Ashrit et al., 2001). Although the climate models simulate increased Indian summer monsoon precipitation due to global warming, the reason may not be simply because monsoon circulation has strengthened but could possibly be because the amount of atmospheric moisture content has increased (Kitoh et al., 1997; Douville et al., 2000a).

It is reported that the MPI model (Ashrit et al., 2001) and the CNRM model (Ashrit et al., 2003) showed no global warming related change in the ENSO–monsoon relationship, although a decadal-scale fluctuation is seen, suggesting that a weakening in the relationship might be within natural variability. Ashrit et al. (2001) further showed that while the impact of La Niña does not change, the influence of El Niño on the monsoon becomes small, suggesting the possibility of
asymmetric behavior in the ENSO–monsoon relationship changes. Figure 17.9 shows long-term variations in the correlation between the JJAS Indian monsoon rainfall and the JJA Niño 3 SST anomalies found through observation and the MRI-CGCM2 simulations (Ashrit et al., 2005). By comparing the model results with observed changes during the late 20th century, it seems that the simulations successfully captured the observed weakening of the teleconnection during that period. Further, the model indicates much weakening in the correlation in the 21st century, particularly after 2050. The MRI-CGCM2 model results support the above hypothesis: due to global warming, the Walker circulation no longer influences India at the time of El Niño because the interannual variability of El Niño is superimposed on an eastward movement of the Pacific circulation system that results from the baseline El Niño-like climate change response.

17.3 WINTER MONSOON

This section briefly reviews a possible change in the Asian winter monsoon due to global warming. Most model results agree that the surface air temperature in wintertime east Asia will rise more than the global average in the future (IPCC, 2001). In winter, the temperature rise over the Eurasian continent is more than that over the oceans, leading to a weakening of north-westerly winter monsoon flow from Siberia toward Japan (JMA, 2000). In particular, a reduction of sea ice in the Okhotsk Sea will bring a remarkable rise in the winter temperature in north-east Asia (Noda et al., 1996; also see Figure 17.3(d)). At northern high latitudes, there is a wavenumber 3 structure in the lower troposphere with troughs over the Okhotsk Sea, the Barents Sea, and Hudson Bay, and the snow line is located closer to the equator than at other longitudes. Since these sea ice areas are located in lower latitudes that receive more solar radiation, the snow–albedo feedback is considered to increase with greater warming.
The time series of the intensity of the northerly wind from the East China Sea to the South China Sea (15°–40°N, 115°–130°E) by the MPI coupled model is shown in Figure 17.10 (Hu et al., 2000a). A weakening in the winter north-easterly winds is seen along the Pacific coast of the Eurasian continent. In the global warming scenario, the contrast between the sea level pressure and the near-surface temperature in the area between the Asian continent and the Pacific Ocean became significantly smaller. The upper troposphere trough and jet stream were found to weaken and shift northward and eastward. The same index by the MRI-CGCM2 is shown in Figure 17.11(a); however, no trend can be seen in this model. When the surface northerly wind in the South China Sea was used as an index (Figure 17.11(b)), the MRI model showed an increasing trend in the latter half of the 21st century. Figure 17.12 shows the distribution of DJF mean surface wind, its magnitude at present, and its change by the end of the 21st century (2071–2100), as determined by the MRI-CGCM2. In the extratropics, there was a northward shift of the cyclone track and the position of the Aleutian Low, which resulted in a weakening of the north-westerly cold surge around Japan. This is similar to the study by Geng and Sugi (2003), which used the T106 JMA AGCM with 20-year time-slice experiments and found a decrease in DJF extratropical cyclone density in east Asia around Japan (south of 50°N) and an increase north of 50°N. Over the South China Sea, on the other hand, the north-easterly wind became strong. This resulted in increased precipitation in the South China Sea and the Indonesian region (Figure 17.3(e)).

Some models showed an increase in winter precipitation around Japan (e.g., Hu et al., 2000a), while others predicted a decrease in precipitation (e.g., Boer et al., 2000); the intermodel difference is large (Giorgi and Francisco, 2000; Lal and Harasawa, 2001; IPCC, 2001), and even the direction of regional change (+ or −) is in dispute when the effect of aerosols is considered (Lal and Harasawa, 2001).
Dai et al. (2001) showed the result of NCAR CSM and PCM. Although the CSM result is noisy, the five-member ensemble mean of PCM shows a 10% to 30% increase in precipitation north of 30°N, and a 10% decrease to the south of 30°N.

As for the north-west monsoon in India, Pal et al. (2001), using the NCAR CCM3 CO₂ doubling experiment, reported an increase in the north-western monsoon precipitation from southern India and Sri Lanka to Indonesia through the Bay of Bengal. This corresponds to a northward shift of the Intertropical Convergence Zone (ITCZ) over the Indian Ocean due to warming. Lal et al. (2001) also reported a decline in the wintertime rainfall in India.

Three regional climate models (MRI, CRIEPI, and NIES) obtained climate change projections for the wintertime Japanese region (Ichikawa, 2004). Sato (2000) double-nested the regional climate models in the Asian region (RSM: horizontal resolution of 120 km) and the Japan region (JSM: 40 km) to the MRI-CGCM1. In a 20-year simulation of the January climate, the precipitation over Japan on the side toward the Japan Sea decreased when there was a decrease in the cold surge from Siberia, while the precipitation on the side toward the Pacific Ocean increased. Kato et al. (2001) performed a 10-year integration by NCAR RegCM2.5 (50 km resolution) nesting to NCAR CSM, and showed that a significant change in precipitation is not seen. On the other hand, in the NIES regional climate
model (50 km resolution), the precipitation from southern China to southern Japan decreased significantly (Emori et al., 2000). Thus, the decreasing precipitation over Japan on the Japan Sea side, which corresponds to weaker cold surges from Siberia, is predicted similarly in the three models, but the models produce different results for the precipitation change over the Pacific side of Japan, suggesting a need to improve the global model.

17.4 OTHER ISSUES

Human beings have changed the land surface through fuel extraction and the expansion of farming areas. The potential (natural) vegetation and the existing vegetation differ greatly from each other in Asia, and further land use changes in
the future are highly probable. Wei and Fu (1998) performed sensitivity experiments looking at land surface change from grassland to desert in northern China, and found that such a land use change will weaken the monsoon circulation and reduce precipitation significantly. Similar results were also obtained when the potential vegetation was altered to reflect the present vegetation (Fu, 2003). Such a vegetation change was actually observed through changes in land surface and land use brought about by past human activity, which suggests the possibility that vegetation changes caused by future anthropogenic land use and climate changes will influence the monsoon circulation and precipitation pattern. Inclusion of the vegetation feedback will produce results showing different regional climate changes (Douville et al., 2000b). There are many GCM experiments that have investigated the impact of deforestation on the climate over various regions. In the south-east Asian monsoon region, Kanae et al. (2001) analyzed the probable precipitation decrease over Thailand due to deforestation over the last 40 years. Sen et al. (2004a) further showed that deforestation on the Indochina peninsula also affects the east Asian rainfall. Chen et al. (2004) and Sen et al. (2004b) discussed the effect of vegetation changes on the regional climate over China.

The influence of global warming on the tropical cyclone climatology, such as its intensity, frequency, and track, is also of great concern. Observed historical changes in some tropical cyclone characteristics have been reported recently (e.g., Yumoto and Matsuura, 2001; Ho et al., 2004). In order to determine future projections of changes in tropical cyclone characteristics, some experiments have used a high-resolution GCM (Bengtsson et al., 1996; Knutson and Tuleya, 2001; Sugi et al., 2002). Contrary to earlier low-resolution model results, they found a significant reduction in the frequency of tropical cyclones with CO₂ doubling. Sugi et al. (2002) showed a regional change in tropical cyclone frequency associated with the SST anomaly, and a significant decrease in typhoons over the North Pacific. The stabilization of the tropical atmosphere due to global warming seems to be responsible for such a change. Yoshimura et al. (2005) found an approximate 20% decrease in tropical cyclone numbers globally. They also showed that mean precipitation near the tropical cyclone centers is significantly larger in the warming experiments than in the present-day climate experiments, as compared with those with the same maximum wind speed. Wu and Wang (2004) estimated the tropical cyclone track changes in the western North Pacific using the Geophysical Fluid Dynamics Laboratory (GFDL) global warming experiments. Based on changes in the large-scale steering flow and in the formation locations in the GFDL model results, they predicted that tropical cyclone tracks will shift eastward in the western North Pacific during the mid-21st century. However, the spatial resolution of these GCMs is limited when representing tropical cyclone structures, and therefore an attempt with a much higher resolution model is needed (Mizuta et al., 2004; Oouchi et al., 2005).
17.5 SUMMARY

Future anthropogenic changes in GHGs, aerosols, and land use/land cover will inevitably alter the climate in Asia. As global warming leads to greater warming over land than over the oceans, the continental land–sea temperature contrast will become larger in the summer and smaller in the winter. Based on this prediction, a simple conclusion is that the Asian summer monsoon will be stronger and the winter monsoon will be weaker in the future. However, model results are not so straightforward and do not fit easily into such a simple view. For the south Asian summer monsoon, models suggest a northward shift of the lower tropospheric monsoon wind system with a weakening of the westerly flow over the northern Indian Ocean. However, atmospheric moisture buildup due to increased temperature will result in larger moisture flux and more precipitation over India. Changes in the winter monsoon will have different consequences from region to region. East Asia will have a weakened north-westerly cold surge, while south-east Asia could experience a stronger north-east monsoon. The projections of these regional climate changes, however, may still be highly model dependent, and should be considered carefully as there are large differences among models in projecting changes in wintertime precipitation, including even the direction of changes.
Figure 17.8. Onset dates of the rainy season. (a) Observation based on Xie and Arkin (1997) precipitation. (b) Present-day (1981–2000) simulation. (c) 2081–2100 in SRES-A2 scenario by the MRI-CGCM2.