11

Land–atmosphere interaction

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11.1 FUNDAMENTAL PROCESSES IN THE SEASONAL CYCLE

The monsoon is manifested as a land–atmosphere–ocean interaction between continents and oceans in the seasonal cycle. The ocean has a large heat content with a longer climate memory of more than a year, but the land has a small heat content and its climate memory is believed to be short (less than a season). Some of the recent model studies emphasized the relative importance of the ocean–atmosphere interaction (OAI) compared with land–atmosphere interaction (LAI), particularly focusing on the strong impact of large-scale sea surface temperature (SST) anomalies in the tropical oceans nearby the continent. It should be noted, however, that to specify or distinguish the roles of LAI from those of OAI and vice versa is very difficult or, particularly in the Asian–Australian monsoon system, because these two processes are strongly coupled to each other. In other words, the LAI (or OAI) in the monsoon system should be understood as part of the full land–atmosphere–ocean interactive system.

The land shows strong and rapid heating (and cooling) in the seasonal cycle which in turn has a large impact on seasonal atmospheric differential heating (and cooling) processes between the land and the ocean. The land surface processes which modulate the seasonal heating, therefore, are likely to be responsible, to some extent, for the interannual variability of the monsoons. The snow cover and soil moisture anomalies over the Eurasian continent in the pre-monsoon seasons have been thought to have a large impact on the Asian summer monsoon variability (Section 11.2).

Seasonal land surface heating and the resultant atmospheric heating over Eurasia manifests itself in the surface or lower tropospheric pressure field. Figure 11.1 shows the difference of the monthly mean geopotential field at 850 hPa from April to May. A remarkable decrease of pressure is seen over southern Eurasia centered over the India/Tibetan Plateau area. Similar patterns of

B. Wang (ed.), *The Asian Monsoon*.

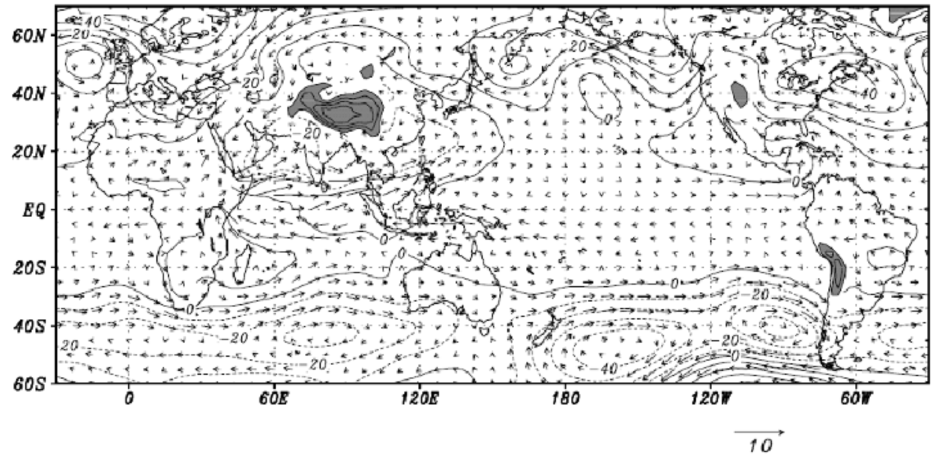

Figure 11.1. Difference of climatological mean height and U, V field for May values minus April values. Dashed line shows negative values.

pressure change are seen over Eurasia from month to month for the period of March through July. In contrast, the pressure change over the tropical and subtropical oceans are very small and positive (increase in pressure). The large decrease of pressure over Eurasia is indicative of the seasonal heating of land, presumably centered over the Tibetan Plateau and Mongolia. The decrease of pressure over North America centered over the Rockies is very small, suggesting smaller heating compared with the Tibetan Plateau area.

The low-pressure area over southern Eurasia (called the 'monsoon trough') at the surface and lower troposphere induces moist south-westerly wind from the Indian Ocean (the south-west monsoon flow) toward India, south-east Asia and east Asia, and dry northerly wind from the interior of the Eurasian continent. The moist monsoon flow, in turn, induces convection and precipitation over the south-eastern part of the continent, which plays a dominant role in atmospheric latent heating as a 'moist LAI' during the monsoon season.

An important issue may be how this moist LAI starts in the interior of the continent. The sensible heating over the Tibetan Plateau may be one of the important factors for triggering this process. In other continents, dry LAI based on sensible heating from the land surface cannot produce the continental-scale monsoons. Figure 11.2 shows a schematic diagram of this dry and shallow LAI. This process is dominant over the Australian continent and west Africa (to the north of the west African monsoon) where desert area prevails and large-scale downward motion dominates. The existence of the desert and the downward motion are self-perpetuating through a positive feedback of surface albedo and radiation balance (Otterman, 1974; Charney, 1975; Webster *et al.*, 1998). This dry LAI cannot induce deep atmospheric heating in the interior of the continents, but can trigger shallow coastal monsoon circulation along the periphery of the continent. This shallow monsoon circulation, however, is likely to trigger deep convection when the

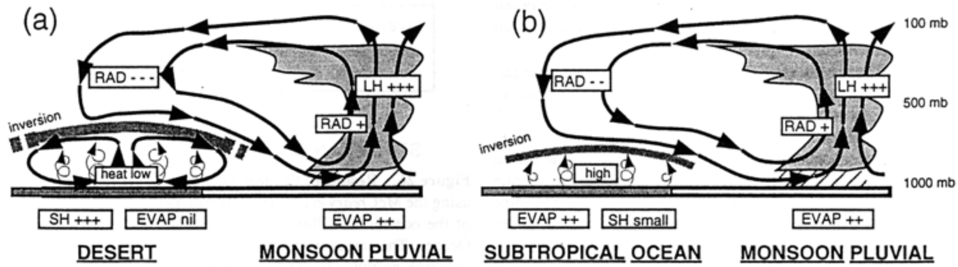

Figure 11.2. Schematic of the circulation between (a) the desert region of North Africa and the Near East; and (b) the subtropical ocean regions and the precipitating (pluvial) part of the monsoon circulation, respectively. The components of the monsoon match the transverse and lateral monsoon components shown in Figure 11.9. The dominant heating or cooling terms are shown in the boxes. 'SH' and 'LH' refer to sensible and latent heating. 'RAD' and 'EVAP' refer to radiational heating and evaporation, respectively. The pluses and minuses represent the sign of the processes and their relative intensities.
Webster *et al.* (1998).

continents are surrounded by warm tropical oceans (e.g., the warm water pool of the Arafura and Timor Seas neighboring the Australian continent) (Kawamura *et al.*, 2002). The land heating process over the Australian continent is thus important for triggering the Indonesian–Australian monsoon.

Over the elevated land surface of the Tibetan Plateau, in contrast, sensible heating plays a dominant role in atmospheric diabatic heating of the whole troposphere particularly in the onset phase of the Asian monsoon. Figure 11.3 (color section) shows diabatic heating rate (Q1) in May over the Tibetan Plateau based upon the GEWEX Asian Monsoon Experiment (GAME) reanalysis data, showing strong heating of the whole troposphere over the Plateau through deep, dry convection (Ueda *et al.*, 2003a). This diagram suggests that latent heating through shallow convection in the pre-monsoon season may also play some role in the heating process even in the preonset phase. The simulation of Asian monsoon by a coupled ocean–atmosphere general circulation model (GCM) also suggests that without the Tibetan Plateau orography the monsoon precipitation cannot penetrate into the interior of the continent (Abe *et al.*, 2003). The role of the Tibetan Plateau is discussed in more detail in Chapter 13.

Diabatic heating over the Tibetan Plateau in the pre-monsoon phase causes horizontal temperature (pressure) gradients in the upper (lower) troposphere between the surrounding Indian and Pacific Oceans and the interior of the continent, which facilitates large-scale convective activity over the South China Sea through the strengthening of the south-west monsoon flow intrusion (Ueda and Yasunari, 1998). Very recently, Jiang *et al.* (2004) and Wang *et al.* (2005) pointed out an important role of the existence of a strong vertical easterly wind shear over south Asia to the south of the Tibetan Plateau in the northward migration of the convective zone with intraseasonal timescales (including the monsoon onset phase) from the equatorial Indian Ocean toward the Himalayas.

This may imply that the atmospheric diabatic heating over and around the Tibetan Plateau (which is primarily responsible for producing the meridional thermal contrast and the vertical easterly shear over south Asia) is one of the essential elements, which characterize the variability as well as the mean state of the Asian summer monsoon system.

11.2 LAND SURFACE QUANTITIES CONTROLLING MONSOONS

The physical quantities of the land surface which may posses anomalous atmospheric forcing or climate memory effects can be (1) snow cover, (2) soil moisture, and (3) vegetation. The land surface layer is generally thought to have a small heat capacity compared with the ocean surface layer, the non-linear processes of these quantities in seasonal and interannual variations effectively produce a relatively longer memory effect, even compatible with SST as shown in Figure 11.4 (Walsh

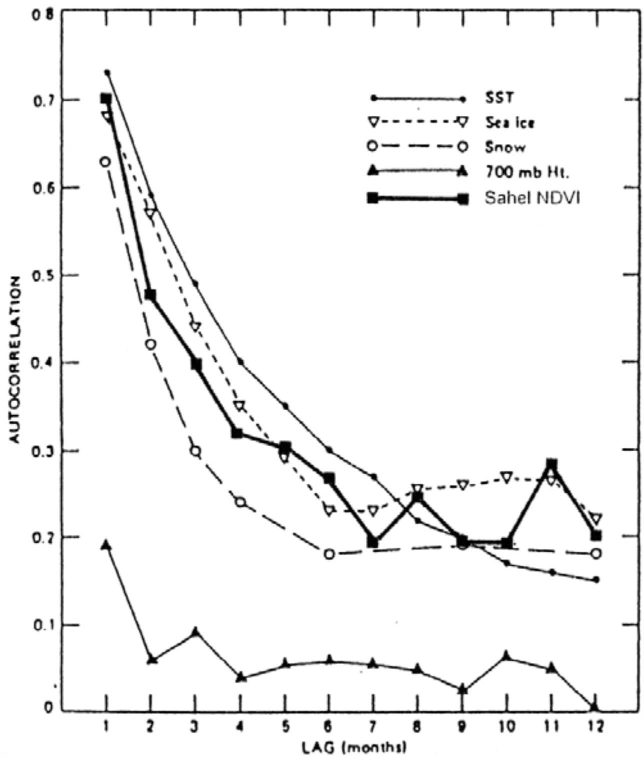

Figure 11.4. Autocorrelations of monthly SST anomalies in the northern Pacific, sea ice in the Arctic Sea, snow cover over Eurasia, geopotential height at 700 hPa depicted from Walsh *et al.* (1985), and NDVI in the Sahel region, North Africa obtained from Shinoda and Gamo (2000). From Shinoda *et al.* (2003).

et al., 1985; Shinoda and Gamo, 2000). The seasonal and interannual variability of these quantities play a significant role in LAI by producing anomalous surface and atmospheric heating or cooling, which in turn affects the monsoon circulation and convection. One should keep in mind that these quantities are interactive with the atmospheric conditions so that detection of a simple cause–result relationship is sometimes difficult. However, due to a lack of adequate observational data, and a lack of understanding of the actual physical processes particularly relevant to climate studies, large uncertainties still exist in the relationship between variabilities of these quantities and the monsoon activity. Comprehensive and interactive observational and modeling strategies are of significant importance to this study.

11.2.1 The role of continental snow cover

Snow cover has several effects which produce anomalous atmospheric conditions as follows: (a) the albedo effect controlling incoming solar radiation, (b) the insulation of heat between the atmosphere and land surface by a snow mass with a small heat conductivity, (c) a heat sink effect resulting from the melting process, and (d) a water source resulting from the melting process (Shinoda *et al.*, 2003). The first three effects can work to change the surface energy transfer process while snow cover exists on the surface. Only the fourth effect possibly has a climate memory even after snowmelt, by affecting the soil moisture content near the surface – called the 'snow–hydrological effect' as shown in Figure 11.5 (Yasunari *et al.*, 1991).

The data sets for validating these effects of snow cover are still limited. Most of the observational studies have been made based on the snow cover extent (of frequency) data from the National Oceanic and Atmospheric Administration (NOAA) operational meteorological satellites. Some studies also used ground-based snow depth measurements for limited areas (e.g., the former USSR, USA, Canada etc.). The continental-scale snow mass (water equivalent depth) information retrieved from satellite-based microwave data have become available over the last one or two decades (e.g., Koike *et al.*, 2001) though some accuracy problems still exist due to biases from vegetation and melting processes.

Blanford (1884) first noted the relationship between the Himalayan winter snow cover anomaly and the succeeding 'all-India' monsoon rainfall (AIMR). Walker (1910) followed up this study and reconfirmed this negative correlation. Since the 1960s satellite-based snow cover extent data became available. Many studies have addressed this issue as a typical example of the role of LAI on the interannual variability of Asian monsoon. Hahn and Shukla (1976) showed an apparent negative correlation between the satellite-based winter snow cover extent anomaly over Eurasia and the following AIMR, though the data used were only for 9 years (Figure 11.6). Since then numerous similar studies have been undertaken using some different indices of snow cover for winter and/or spring, and some different indices of the Indian or Asian summer monsoon activity (Dey and Bhanu Kumar, 1982; Dickson, 1984; Bhanu Kumar, 1987, 1988; Chattopadhay and Singh, 1995; Kripalani *et al.*, 1996; Shankar-Rao *et al.*, 1996; Morinaga *et al.*, 1997, 2000; Bamzai and Shukla, 1999; Kripalani and Kulkarni, 1999; Kripalani *et al.* 2003;

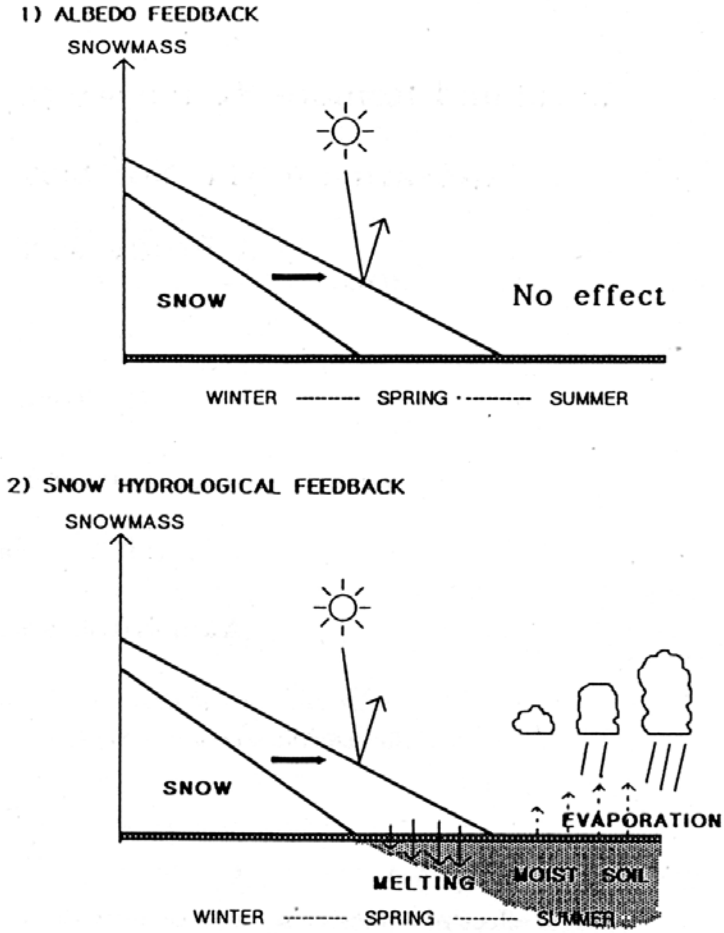

Figure 11.5. Schematic diagram for the albedo feedback and the hydrological feedback of snow cover during the seasonal march from winter to summer.
Yasunari *et al.* (1991).

Robock *et al.*, 2003). Some studies also discussed the Eurasian snow–AIMR relationship relating to the El Niño/Southern Oscillation (ENSO)–monsoon relationship (Yasunari, 1987; Kahndekar, 1991; Yasunari and Seki, 1992; Yang, 1996; Yang and Lau, 1998; Kawamura, 1998; Ye and Bao, 2001; Liu and Yanai, 2002). Some of these studies concluded that the snow cover–AIMR correlation might result from a common large-scale atmospheric circulation pattern that is responsible both for the snow cover anomaly and the AIMR anomaly. One possible idea is that ENSO-related teleconnection induces anomalous atmospheric circulation over Eurasia, which may fundamentally be responsible for anomalous Asian summer monsoon activity (Webster *et al.*, 1998). The snow cover anomaly induced by the anomalous circulation may be partly reinforced to produce a weaker monsoon

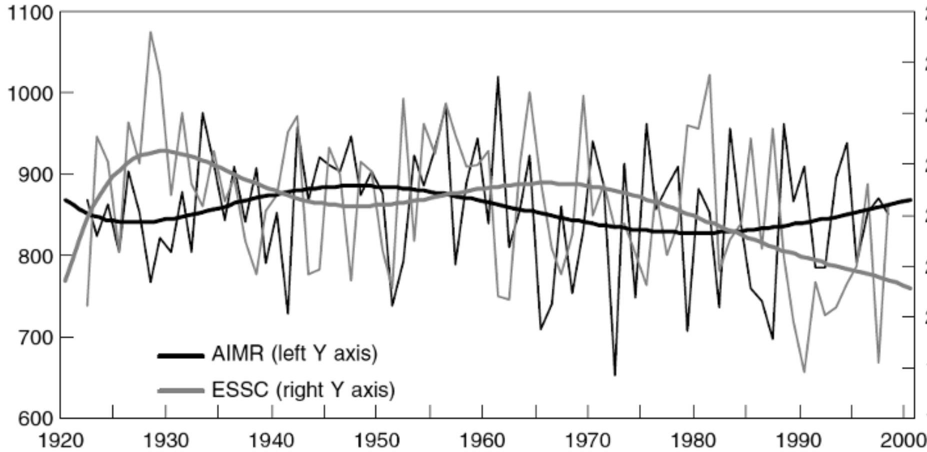

Figure 11.6. Year-to-year variations of the AIMR and ESSC. The thick solid lines represent the trends fitted with sixth-order polynomials.
Liu and Yanai (2002).

through radiation and energy flux processes as suggested in the observational study of Yang *et al.* (2004) and in a GCM experiment of Shen *et al.* (1998).

A central issue for the snow cover–AIMR connection may be how the climate memory of the snow cover anomaly in winter (or spring) can be conveyed to the following summer. Some GCM studies attempted to address this issue (Barnett *et al.*, 1989; Yamazaki, 1989; Yasunari *et al.*, 1991; Vernekar *et al.*, 1995; Douville and Royer, 1996). Though some of these GCM experiments have given unrealistically large anomalous snow covers (snow mass) as an initial condition, both the albedo and the snow–hydrological effect of anomalous snow cover could play, to some extent, the role of producing monsoon anomalies in the summer. Some studies emphasized that the albedo and snow-melting effect on the Tibetan Plateau in spring showed a relatively large impact on weakening of the summer monsoon through suppressing the net radiation and sensible heat flux (Yasunari, 1991; Douvelle and Royer, 1996). In contrast, a recent large-ensemble GCM experiment has suggested that in the northern part of the continent (e.g., Siberia), snow mass anomaly may be more important for producing an atmospheric anomaly than snow cover (albedo) forcing (Gong *et al.*, 2004). Takata and Kimoto (2000) pointed out the importance of seasonal freezing (and melting) of the surface permafrost layer in the northern part of the continent on thermal and hydrological processes of the surface. Full inclusion of these processes causes higher surface temperatures (due to suppressed evaporation) and stronger monsoon circulation in south-east Asia.

The recent precise observational studies on the seasonal march of temperature and circulation fields over Eurasia from spring to summer have shown that the influence of snow cover and related soil moisture anomalies on the temperature and circulation anomalies in the lower troposphere is limited primarily when and where snow cover exists seasonally as schematically shown in Figure 11.7

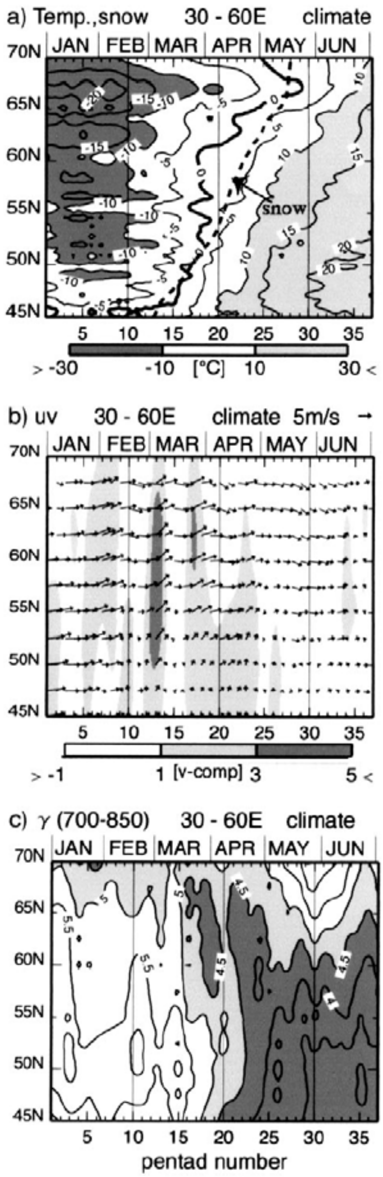

Figure 11.7. Latitude–time sections showing (a) mean (1966–1990) seasonal evolution of the surface temperature (°C) along a longitude of 30°–60°E. Thick black contour indicates a surface temperature of 0°C. The dashed contour denotes climatological snow disappearance pentads. (b) The same as in (a), except for 850 hPa horizontal wind obtained from ECMWF reanalysis (1980–1990). The southerly wind component greater than $1\,\mathrm{m\,s^{-1}}$ is denoted by two-tone shading. (c) The same as in (a), but for static stability γ between 850 and 700 hPa. Shading represents the region of γ less than $5.0\,\mathrm{K\,km^{-1}}$.
Ueda *et al.* (2003b).

(Shinoda *et al.*, 2003; Ueda *et al.*, 2003; Robock *et al.*, 2003). The long-term surface energy flux observation on the Tibetan Plateau as part of GAME in 1998 suggested that the anomalous heavy snow cover in spring on the central Plateau resulted in a considerable decrease of sensible heating and delayed the monsoon onset over the Plateau (Tanaka *et al.*, 2004). The observational study (Ohta *et al.*, 2001) of seasonal surface energy and water flux in east Siberia conducted as part of GAME–Siberia also proved the important processes noted by Takata and Kimoto (2000).

11.2.2 The role of soil moisture

Soil moisture is a key parameter for LAI in the climate system. The anomaly of surface and near-surface soil moisture is likely to have a persistence of several days to several months, which may cause climate memory through anomalous surface energy and moisture fluxes. However, this quantity is very difficult to measure adequately. The long-term, *in situ* measurement of soil moisture is very limited both in time and space (Robock *et al.*, 2000, 2003). The recent satellite-based indirect measurement using microwave sensors is a promising method for large-scale soil moisture monitoring, but it has still been limited for soil moisture or wetness of the very near surface soil layer. The climate memory effect of soil moisture has, therefore, been discussed basically based upon climate model experiments.

Delworth and Manabe (1988, 1989) firstly assessed the climate memory effect of soil moisture in GCMs using the so-called bucket model (Manabe, 1969) interacting with the atmosphere through a surface water balance. Their bucket model consisted of one soil layer with a 15-cm water holding capacity (field capacity). They noticed that the persistency (measured by autocorrelation) of the soil moisture anomaly was small whereas in higher latitudes its value was large, depending upon the ratio of P/Ep (P, precipitation; Ep, potential evapotranspiration) as shown in Figure 11.8. In the humid tropics or monsoon regions where P is large enough to saturate the soil layer, the soil moisture anomaly does not function as a climate memory. The distribution of persistency in Figure 11.8 can be generally described in the form $1/\lambda = W^*/Ep$, where λ is e-folding time of exponential decay of soil moisture anomaly, and W^* the field capacity. Though their definition of soil moisture and assumption of soil layer model were so simple, this relation between soil moisture anomaly persistency, field capacity of soil layer, and Ep was partly proved in observations in central Eurasia (Vinnikov and Yeserkepova, 1991).

In recent years, more realistic and sophisticated land surface models (LSMs) have been developed, and many sensitivity experiments with GCMs have been conducted using different types of LSMs under different experimental designs. For example, Douville *et al.* (2001) and Douville (2002) assessed the influence of soil moisture on seasonal and interannual variability of the Asian and African monsoons. Through some ensemble experiments with and without an interactive LSM (named ISBA; Interaction between Soil Biosphere and Atmosphere model), they found that the impact of soil moisture anomaly through the LAI process is significant in relatively dry monsoon regions (e.g., the Indian subcontinent and

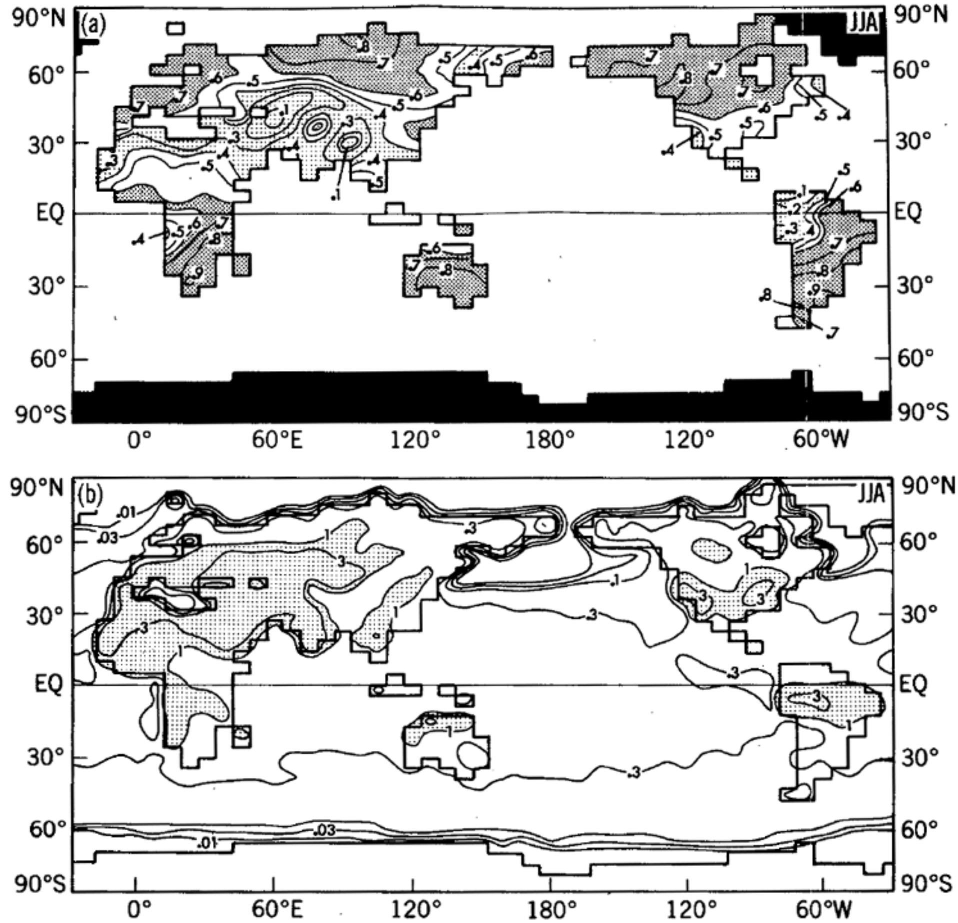

Figure 11.8. (a) Lag one autocorrelation values of soil moisture for the months of June, July, and August (JJA) for SMI. At each grid point, deviations of monthly mean soil moisture from the long-term mean for that month were correlated with data from the same grid point, but lagged by one month. Coefficients greater than 0.16 (0.3) are significantly different from zero at the 95% (99.9%) confidence level (see Chatfield, 1984, p. 63). Values greater than 0.6 are densely stippled, while values less than 0.4 are lightly stippled. Permanently ice covered regions are black. (b) Potential evaporation (cm d^{-1}) for JJA in SMI.
From Delworth and Manabe (1989).

north-east Asia) but not in the humid monsoon regions (e.g., south-east Asia). They also noticed that in the significantly impacted regions soil moisture anomaly primarily plays a role as an initial condition (at the monsoon onset phase) rather than the boundary condition. Kanae *et al.* (2004) also performed a set of hindcast GCM simulations (with prescribed SST) on the boreal summer hydroclimate for about 40 years (1951–1998), with and without realistic soil moisture anomalies, and noticed that only semiarid regions in the peripheries of monsoon regions showed reasonably good simulated precipitation compared with that observed. It is interesting to note that all the GCM experiments, though the LCM performance and simulation design are quite different from each other, have shown a relatively high impact of soil moisture anomaly in dry or semiarid regions. This general tendency of soil–precipitation feedback has comprehensively been confirmed by the recent Global Land Atmosphere Coupling Experiment (GLACE) (Koster *et al.*, 2004), where twelve GCMs with different LCMs participated. For example, in the Asian monsoon region, the sensitivity of precipitation change to soil moisture condition is large only in relatively dry areas (e.g., the Indian subcontinent and the Yellow River/inner Mongolia in China) as shown in Figure 11.9 (color section). In humid monsoon and tropical regions, the impact seems to be small or insignificant primarily due to already saturated soil moisture conditions. However, this characteristic nature of the weak soil moisture memory effect in the humid tropics may, to some extent, counteract with the relatively long memory effect in the humid tropics (Wu and Dickinson, 2004). In addition, the soil moisture anomaly may be important as an initial condition to switch on the interaction with precipitation in these relatively dry regions.

11.2.3 The role of vegetation and landuse/land cover changes

Some new aspects from observational studies

Another important aspect of LAI in addition to soil moisture (and snow cover) may be the role of vegetation. In the Asian monsoon region, most of the areas are heavily covered by vegetation, including tropical rain and monsoon forests, water-fed paddy fields, grass lands, and boreal forests, etc. The roles of vegetation in LAI may be classified as follows:

(1) control of radiation and energy fluxes through albedo and surface roughness;
(2) control of transpiration through stomatal resistance; and
(3) control of the substantial field capacity of soil with root depth and structure.

Due to a lack of observations particularly of the large-scale vegetation and its impact on climate and the water cycle, this aspect has been noted only in the recent years. LSMs including these vegetation processes have also been developed and improved in the most recent few decades (e.g., Dickinson and Henderson-Sellers, 1988; Sellers *et al.*, 1986), though these models need numerous numbers of tuning parameters. To improve and fully utilize these sophisticated models, however, observational studies are also essential, including optimal determination of these parameter values. In the Asian monsoon region, intensive field campaigns related to some international projects such as GAME have been conducted, and have revealed new aspects of the role of vegetation in terms of energy and water cycle processes.

In the evergreen tropical forest in Thailand, energy and water flux measurements at several sites with different types of vegetation were conducted for more than two years as part of GAME–Tropics. The evapotranspiration estimated by a multilayer

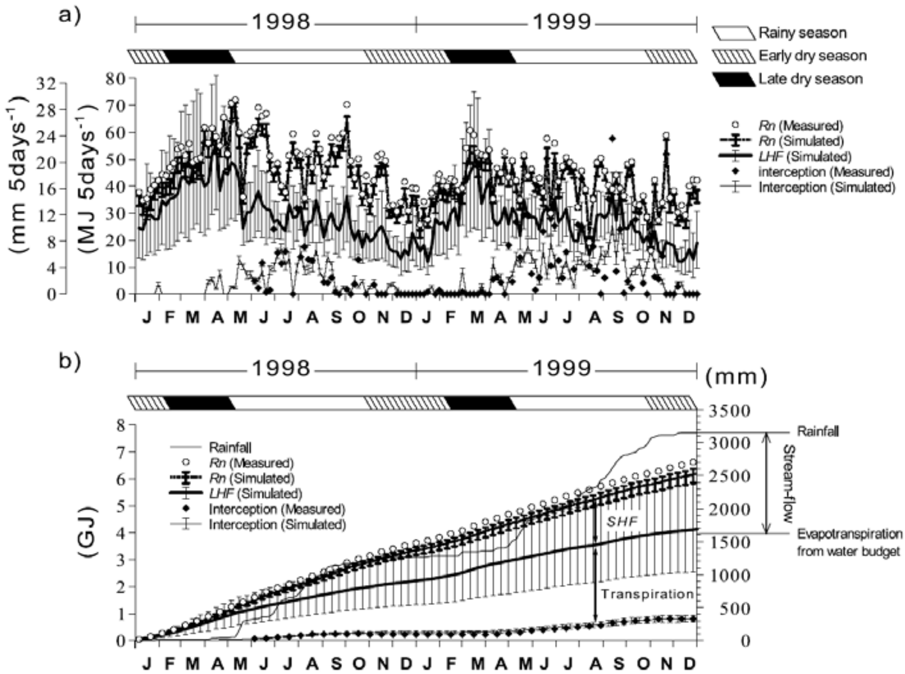

Figure 11.10. (a) Simulation of the seasonal variation in the net radiation and latent heat over an evergreen forest and rainfall interception in 1998 and 1999. (b) The cumulative result. In the figure the tops and bottoms of the vertical bars indicates the maxima and minima, respectively, of the 32 simulated values. The line in between shows the average.
Tanaka *et al.* (2003).

model forced by the observed radiation, leaf area index (LAI), and other realistic parameters revealed that a seasonally maximum evapotranspiration appeared in the late dry season (Figure 11.10) just before the monsoon onset when the ground surface was at its driest for the year (Tanaka *et al.*, 2003). An LSM study including the effects of root depth (to pump up the water in the deep soil layer) was found to simulate this seasonal feature well when the mean root depth was set to 6 m (Tanaka *et al.*, 2004). This implies that effective field capacity must be set at an extremely large value in this case. In the deciduous monsoon forest in Thailand, in contrast, the seasonal maximum of evapotranspiration appeared in the late monsoon season, associated with a seasonal photosynthetic activity and water availability (Toda *et al.*, 2002). These dependencies of the seasonal variation of evapotranspiration (i.e., latent heating) on dominant vegetation types may have a great impact on the modeling of the regional and large-scale climate and water cycle, particularly in the tropics.

In the boreal forests of Siberia, similar long-term flux measurements were made in the Lena River basin as part of GAME–Siberia (Ohta *et al.*, 2001) as shown in Figure 11.11. Here, surface energy partition (sensible vs. latent energy) is strongly

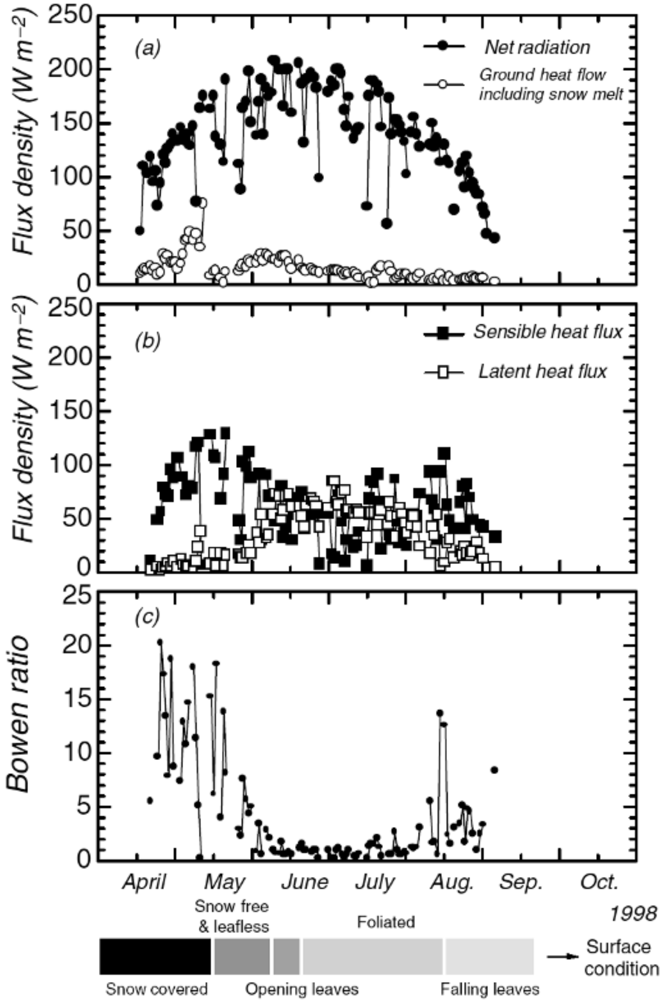

Figure 11.11. Time series for the daily net 'all-wave' radiation and the ground heat flow (a), the sensible heat flux and the latent heat flux (b), and the Bowen ratio (c). The energy balance components and the Bowen ratio indicated in this figure are for a dry canopy only.
Ohta *et al*. (2001).

controlled by the phenology of the (larch) forest coupled with seasonal melting (and freezing) conditions of the permafrost beneath the forest. Namely, in May during and after the snowmelt, sensible heating greatly increased, which was mostly used for melting the frozen ground. At the beginning of June when the melt layer of the ground surface reached about 20 cm depth or so, the foliation of trees suddenly started, which at the same time triggered transpiration (latent heating) from the trees. During mid-summer (June, July and the beginning of August) latent heating rate was comparable with, or more than, the sensible heating rate. Most of the annual precipitation occured in this period. This condition prevailed for a broad area of the Siberian boreal forest zone, which contributes to water recycling as well as maintenance of permafrost, by limiting the melting of permafrost within the near-surface layer. Thus, the boreal forest and permafrost of this region are likely to coexist as a symbiotic ecoclimate system through energy and water cycles.

Some new aspects from modeling studies

The potential role of large-scale vegetation and land cover on the Asian monsoon has recently also been investigated by some GCM experiments. Xue *et al.* (2004) showed that some characteristic features of the east Asian summer monsoon (e.g., the abrupt northward jump of the rainfall belt in the seasonal march and low-level monsoon circulation) are far better simulated when a more sophisticated vegetation scheme is introduced (a simplified simple biosphere (SSiB) with realistic vegetation distribution) rather than a simple vegetation scheme without physiological processes (i.e., photosynthesis and stomatal control of transpiration). The abrupt northward jump of rainfall could occur due to strong sensible heating in the interior of east Asia which would induce horizontal temperature and pressure gradients in the pre-monsoon season. This would probably relate to stomatal control of evapotranspiration in the dry and hot season, resulting in enhancement of sensible heating. Suh and Lee (2004) also showed in the regional climate model experiment that temperature and rainfall biases over east Asia have improved considerably as a result of including the biospheric processes of realistic vegetation over the east Asian land mass. These experiments have suggested that physiological control of surface energy flux partitioning (i.e., Bowen ratio) greatly affects the pressure gradients over land, and in turn, circulation and rain systems. However, the reality of these biospheric processes needs to be validated through *in situ* and satellite-based comprehensive observations.

Very recently, we have conducted a series of GCM experiments assessing the role of the continental land surface with and without vegetation, and relative importance of topography and vegetation on the formation of the Asian summer monsoon. In this case, various vegetation types are simply represented by differences in albedo, roughness, and field capacities of surface soil layers (Yasunari *et al.*, 2005; Saito *et al.*, 2005). These experiments do not explicitly include the physiological processes of vegetation in the model, but the difference in field capacity of the soil layer is expected to simulate the important nature of vegetated or non-vegetated surfaces since the soil layer is basically formed by plants. The results suggest the role of vegetation is generally significant in producing a strong monsoon, and relative roles of albedo and field capacities of soil appear to be different from region to region. In east Asia both the Tibetan Plateau's topography and vegetation are important for the penetration of the moist monsoon flow and precipitation. In addition, the relative importance of albedo vs. field capacity seems to be different from region to region. In east Asia the albedo effect of vegetation is more important, but in south/south-east Asia the role of a large field capacity has proved to be more important in producing stronger monsoons particularly in the mid-monsoon season,

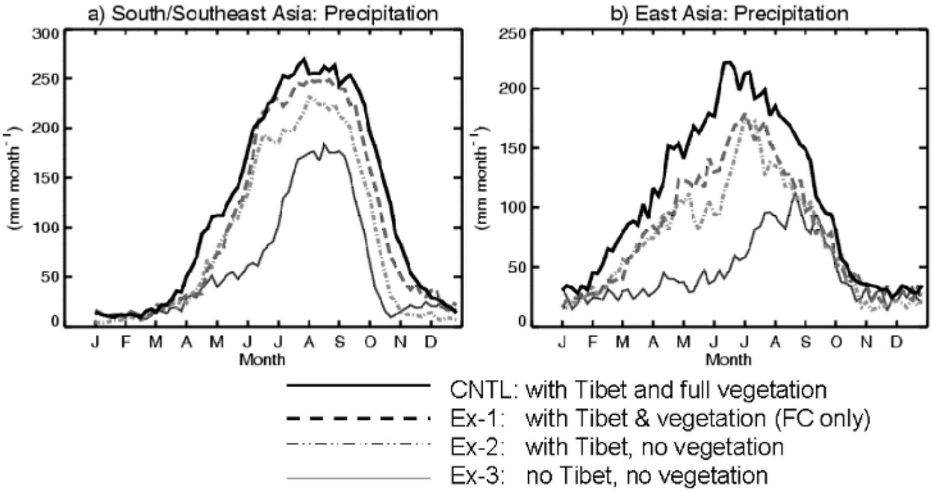

Figure 11.12. Changes of seasonal precipitation for (a) south/south-east Asia and (b) east Asia for different boundary conditions: (1) with no Tibetan Plateau, no vegetation (bared surface with rock albedo and roughness) (thin solid line); (2) with the Tibetan Plateau, but no vegetation (dotted and dashed line); (3) with the Tibetan Plateau and vegetation (soil only) (thick dashed line); and (4) with the Tibetan Plateau and full vegetation (with soil and realistic vegetation albedo and roughness) (thick solid line).
From Yasunari *et al.* (2005).

as shown in Figure 11.12. These numerical experiments have strongly suggested that in the Asian monsoon region the role of vegetation is essential for the formation of monsoon circulation and precipitation through strong latent heating of the atmosphere. The evapotranspiration from vegetated surfaces plays a crucial role in forming a thick moist boundary layer and convection for latent heating, as will be described in the next section.

11.3 THE CONNECTION BETWEEN THE LAND SURFACE, THE ATMOSPHERIC BOUNDARY LAYER, AND CLOUD/PRECIPITATION PROCESSES

The Impacts of land surface processes on the atmosphere are brought about through the modification of the atmospheric boundary layer (ABL), and relevant cloud processes at or above the ABL. Betts *et al.* (1998) reviewed the overall possible interactive processes between land surface conditions and the ABL, and emphasized the importance of vegetation and related land cover/use differences on the ABL structure and cloud/precipitation process. In the monsoon region, the advection or inflow of moist air from oceans to the land is sometimes emphasized to produce monsoon rainfall in the interior of the land area. However, through some continental-scale experiments (CSEs) under GEWEX (e.g., GAME in Asian monsoon region, LBA in Amazon basin) moisture supply from wet land surfaces has also been proved to be important particularly in the late monsoon season when the surface becomes wetter than in the earlier stage of the monsoon season. Shinoda and Uyeda (2002) examined a possible impact of water-fed paddy fields in east Asia, using a cloud-resolving atmospheric model. They forced the model using latent and sensible heat fluxes over the paddy field observed in GAME–Huaihe Basin Experiment (HUBEX), and found that mesoscale cloud systems embedded in the Meiyu (Baiu) front developed through the moisture supply to the ABL and the change of the moist static stability of the lower troposphere. This change of LAI and its impact on cloud/precipitation systems through the seasonal change of soil moisture conditions was also observed over the Tibetan Plateau during GAME–Tibet (Yamada and Uyeda, 2004). In fact, these characteristic changes of the water balance or the recycling process of water vapor between land surface and atmosphere through the seasonal march of the summer monsoon have been noticed in various parts of the continent (Yasunari and Kozawa, 2005). Understanding of these non-linear feedbacks from precipitation to land surface conditions may have some clue for the prediction of seasonal as well as interannual variations of precipitation in the Asian monsoon region.

11.4 THE POSSIBLE IMPACT OF ANTHROPOGENIC LANDUSE/LAND COVER CHANGES ON THE ASIAN MONSOON CLIMATE

The Asian monsoon region is heavily populated with nearly 60% of the World's population. Therefore, the anthropogenic landuse and land cover changes have been significant. These landuse/land cover changes from the original vegetation have the significant possibility to change the regional climate and water cycle. Fu (2003) investigated the potential impacts of human-induced land cover change on the east Asian monsoon, assuming the present and the potential vegetation (Ojima, 2000). Using a regional model (MM5) with a land surface scheme (BATS), he noticed that the land cover change which occurred over the history of China may have weakened the east Asian summer monsoon and strengthened the winter monsoon. The weakened monsoon trough in the interior of the continent presumably induced by a decreased latent heating is responsible for the weakened summer monsoon.

Xue (1996) focused on the impact of desertification in the Mongolian and the inner Mongolian grassland on the east Asian monsoon climate, by changing the area of desertification in the GCM. He found that the desertification is further intensified by altering from the grassland-type to a desert-type land surface due to the reduction in evaporation and convective latent heating above the surface layer. This process seems to be contrastive to the positive feedback of desertification in the Sahel, west Africa where enhanced radiative cooling and an associated sinking motion induced by increased albedo is likely to be a main mechanism (Charney, 1975).

In the Indo-China peninsula, a decreasing trend of monsoon rainfall has remarkably only been seen in September during the past several decades. Kanae *et al*. (2001) tried to explain this feature as a result of deforestation in the central part of the peninsula using a regional climate model (RAMS). They noticed that the effect of deforestation by changing albedo, roughness, and soil moisture conditions significantly reduced rainfall only in September when the monsoon westerly flow seasonally became weak as shown in Figure 11.13 (color section). In other monsoon months, the effects of deforestation were negligible because the effect of a strong inflow and convergence of the moist monsoon flow dominated over this region. These results suggested that the impact of land surface conditions (i.e., vegetation cover, soil moisture etc.) change depending upon the atmospheric conditions including large-scale wind fields, thermal stability, etc.

11.5 THE FEEDBACK PROCESSES THROUGH THE ENERGY AND WATER CYCLES

A key process of regional climate and water cycle change by the landuse/land cover change including deforestation may be a process by which the land surface change induces changes of moisture convergence (C) and *in situ* evapotranspiration (E) which results in a change in precipitation (P). This issue can be more generally interpreted as a fundamental energy and water cycle process of an interaction between an area-limited land surface condition and large-scale atmospheric environment or circulation. This issue was discussed in terms of deforestation in the Amazon basin by Zeng *et al*. (1996) and Zeng (1998). Figure 11.14 shows a schematic diagram of the two feedback loops in the perturbed land surface area for the atmospheric water cycle. Change in P is controlled by two mechanisms: C feedback (right-hand side) and E feedback (left-hand side). In the first mechanism, an increase in P releases latent heat that drives large-scale upward motion (w), which causes more moisture convergence, leading to more P. In the E feedback, higher P leads to a

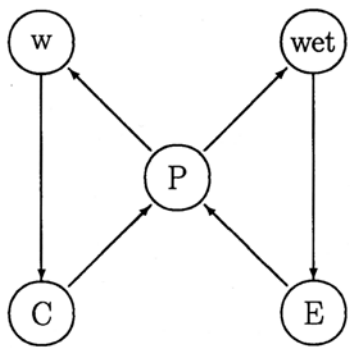

Figure 11.14. The two major feedback loops in the perturbed region: moisture convergence feedback and evaporation feedback. Here, P is precipitation, w is large-scale upward motion, C is large-scale moisture convergence, wet is surface wetness, and E is evapotranspiration.
From Zeng (1998).

wetter surface and more evaporation, which in turn contributes to even more P. If these two positive feedbacks overlap and work at the same time, a higher sensitivity of P to land surface changes appears. However, if the P increase causes a decrease in radiation through, for example, cloud cover increase, the increase of C is partially compensated by a decrease of E due to decrease of radiation. An essential issue in the C feedback may be how a P increase contributes to an increase of large-scale upward motion (w).

Assuming thermodynamic balance and neglecting horizontal temperature advection, radiative heating approximated by Newtonian cooling, convective latent heating proportional to P, the moisture closure (i.e., $P = E + C$), and C proportional to w and humidity q in the boundary layer, the P change (ΔP) can be approximated as a simple equation (Zeng *et al.*, 1996; Zeng, 1998):

$$\Delta P = q\eta \Delta T^{*}/(1 - q - \beta) \tag{11.1}$$

where q is a measure of moist static stability (unstable if $q > 1$), T^{*} is an atmospheric equilibrium temperature, $\beta (\equiv Ep/P_0)$ is the ratio of potential evaporation Ep to a threshold precipitation rate P_0, above which soil is saturated, and η is a constant relevant to the radiation relaxation and atmospheric latent heat profile.

In the tropics near the equator, this positive C feedback roughly depends upon whether the horizontal scale of the perturbed area is larger than the equatorial radius of deformation (Zeng, 1998). In the off-equatorial monsoon region, however, the influence of large-scale moisture advection is dominated and the feedback loops (as shown in Figure 11.14) may not be so simple. The impact of deforestation may be apparent through the two positive feedback loops when the environmental monsoon flow is weak, as demonstrated in the September case in Thailand (Kanae *et al.*, 2001). In the relatively dry or marginal monsoon region (e.g., the north-west part of the Indian subcontinent, Mongolia, and northern China) where the sensitivity of P to the soil moisture anomaly is large (as discussed in Section 11.2.2), the positive feedback loops as shown in Figure 11.14 may be strong mostly due to large β under relatively dry conditions. We should keep in mind that relatively humid regions under the Asian summer monsoon climate, the ABL, and the lower troposphere are close to saturated (i.e., q is relatively large). This condition also increases the sensitivity of P to slight changes of surface conditions (e.g., albedo) and in turn changes of radiational forcing. The large impact of water-fed paddy fields in China on the development of cloud/precipitation systems in the Meiyu frontal zone may correspond to this situation.

This simple model of energy and water balance can be applied to understanding the role of the LAI of the large-scale Asian monsoon system. For example, when we compare the pre-onset phase and the mature phase of the monsoon season, we notice that ΔP between the two season is basically due to the change of T^{*} derived from the seasonal change of solar radiation, but the enhancement of ΔP is due to a change of q and β, derived from moisture convergence and *in situ* evaporation, respectively (Yasunari and Kozawa, 2005). Through these moist processes, the large-scale Asian monsoon system, with a huge P particularly over the land area, could be formed as it is now.

11.6 CONCLUDING REMARKS

This chapter has discussed the role of LAI in the Asian monsoon system. Particular emphasis and focus has been placed on what is the essential and primary role of the 'land' of the Asian (or Eurasian) continent in the Asian monsoon system as a huge coupled ocean–land–atmosphere system.

The role of the Tibetan Plateau is emphasized as a means of triggering the atmospheric heating over the continent. At the onset or pre-monsoon phase of the monsoon, sensible heating over the elevated land surface plays an important role in forming the continental-scale heat low in the lower troposphere. This circulation system in turn produces a moist south-westerly flow penetrating into south, south-east, and east Asia. This moist monsoon flow induces convection and precipitation over the south-eastern part of the continent, which further intensifies atmospheric diabatic heating through latent heat release.

As major land surface parameters controlling the Asian monsoon variability, we have discussed the roles of snow cover, soil moisture, and vegetation. The snow cover and soil moisture are internal factors of the climate system particularly in the seasonal to interannual timescales, and are likely to play some important roles in changing the large-scale surface energy and water balance, which in turn affects the monsoon circulation and precipitation. One big problem is that the data available on these two quantities as internal parameters of the climate system is still limited, though new satellite data are considerably improving this situation. The quantitative estimates of the impact of these parameters on the monsoon-related fields needs further study, but it may be concluded that both the extent of these parameters and what role they play in the variability of the monsoon system depends strongly on the basic climatic conditions and seasonality.

Another parameter, vegetation, has been noted as an important variable for the formation of moist monsoon flow over the continent. Vegetation has several functions for controlling atmospheric energy and water vapor conditions, such as albedo, roughness, stomatal conductance, and water-holding capacity (of the root/soil structure). In fact, through the recent observations and modeling studies, the atmospheric latent heating over land has been noticed fundamentally through vegetation control of evapotranspiration. The anthropogenically induced change of land cover/landuse, including deforestation, has had a great impact on the regional precipitation and water cycle of the monsoon region by changing some characteristics of the vegetation's control of the energy and water cycles, which in turn affect the ABL and cloud/precipitation processes on regional-scales.

Finally, a general discussion has been made on how the land surface changes could induce changes in precipitation through the feedback processes of moisture convergence and *in situ* evapotranspiration processes. In this simple discussion the critical role of moisture amount near the surface and in the ABL is emphasized, to induce a positive feedback to change precipitation over land in the monsoon region.

In the Asian monsoon system, the fundamental heating centers are located over the warm oceans near the continent, which also prove to result from the strong thermal and thermodynamic effect of the Tibetan Plateau (Kitoh, 2004; Abe *et al.*, 2004). In addition, we emphasize that the moist land surface could play an important role for the penetration of precipitation into the deep interior of the continent. This moist land surface process has proved to be essentially attributed to the existence of vegetation. In this sense, the heavily vegetated land surface may be another key factor of the strong Asian summer monsoon in addition to the Tibetan Plateau.