19. LAND-ATMOSPHERE-INTERACTION

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1. Monsoons as Land-Atmosphere-Ocean Interaction

The monsoon is manifested as an atmospheric circulation system of land-atmosphere-ocean interaction (LAOI) between continents and oceans in the seasonal cycle. The ocean has a large heat content with 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 (of less than a season). However, the land shows strong and rapid heating (and cooling) in the seasonal cycle which in turn has a large impact on seasonal atmospheric heating (and cooling) processes. The land surface processes which modulate the seasonal heating, therefore, should also be responsible for interannual variability of the monsoons.

Some of the recent model studies emphasized relative importance of the ocean-atmosphere interaction (OAI) compared to 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, may be nonsense, particularly, in the Asian-Australian monsoon system because these two processes are strongly coupled to each other. In other words, the LAI (or AOI) in the monsoon system should, in any case, be understood as part of LAOI.

In this review report, we will discuss the roles of LAI in the seasonal cycle and interannual variability of the Asian and Australian summer monsoon system, based on the recent observational and modeling studies.

2. Land Surface Quantities in LAI

2.1 Persistencies of Land Surface Quantities

The physical quantities of land surface which may posses anomalous atmospheric forcing or climate memory effect can be 1) snowcover, 2) soil moisture, and 3) vegetation. The snowcover has several effects on atmosphere and surface soil layer as follows: a) albedo effect controlling incoming solar radiation, b) insulation of heat between the atmosphere and land surface by snowmass of small heat conductivity, c) heat sink effect by melting process, and d) water source by melting process. The first three effects can work to change surface energy transfer process while snowcover exists on the surface. Only the fourth effect possibly has a climate memory even after snowmelt, by affecting soil moisture content near the surface which is called "snow-hydrological effect" (Yasunari *et al.* 1991). The snowcover and soil moisture anomalies over the Eurasian continent in the pre-monsoon seasons have been thought to have large impact on the interannual variability of Asian summer monsoon.

Walsh *et al.*(1985) and Shinoda and Gamo (2000) investigated persistencies of some ocean and land surface parameters as shown in Fig. 1, and noticed that the land surface parameters (snowcover, NDVI) also have climate memory effect comparable to ocean surface parameters of sea surface temperature (SST) and sea ice extent. Delworth and Manabe (1988, 1989) investigated time-space characteristics of persistency of soil moisture, by using a GCM with a simple bucket-type land surface hydrological model. They showed a general tendency of long persistency (of more than 0.6 of one-month auto correlation) in cold and wet regions and short persistency (of less than 0.4) in warm and dry regions. The validity of this space characteristics were partly validated by observed soil moisture data (Vinnikov and Yeserkepova, 1991; Vinnikov *et al.*, 1996, 1999, Entin *et al.*, 2000), suggesting that the time-scale of soil moisture anomalies (defined with 1/e limit of the autocorrelation) may be at most 2-3 months.

2.2 Data Sets for Land Surface Quantities

The data sets for validating these effects of snowcover are still limited. Most of the observational studies have been made based on the snowcover extent (of frequency) data from the NOAA operational meteorological satellites. Some studies also used ground-based snow depth measurement 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 has become available for the recent one or two decades (e.g., Koike *et al.*, 2001) though some accuracy problems still exist due to biases from vegetation and melting processes.

The soil moisture is a key parameter for land-atmosphere interaction in the climate system. The snowcover and vegetation index (NDVI) are closely related to soil moisture as discussed later. The anomaly of surface and near-surface soil moisture generally is likely to have 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), and the recent satellite-based indirect measurement using micro-wave sensors has still been limited for soil moisture or wetness of very-near surface soil layer. The climate memory effect of soil moisture has, therefore, been discussed basically based upon climate model experiments.

3. Role of LAI in the Seasonal Cycle of Asian Summer Monsoon

3.1 Seasonal Change of Circulation over Eurasia

Seasonal land surface heating and resultant atmospheric heating over Eurasia is manifested from surface or lower tropospheric pressure field. Fig. 2 shows difference of monthly mean geopotential field at 850hPa from April to May. Remarkable decrease of pressure is seen over southern Eurasia centered over India/Tibetan plateau area. A similar pattern of pressure change is seen over Eurasia from month to month for the period of March through July. In contrast, the pressure change over the tropical and sub-tropical oceans is very small and positive (increase of pressure). The large decrease of pressure change over Eurasia indicates seasonal heating of land, presumably centered over the Tibetan Plateau and Mongolia. The decrease of pressure over north America is very small centered over the Rockies, suggesting smaller heating over there compared to the Tibetan Plateau area.

The low-pressure area (called the "monsoon trough") over the southern Eurasia at the surface and lower troposphere induces moist southwesterly wind from the Indian Ocean (the southwest monsoon flow) toward India, southeast 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 of atmospheric heating during the monsoon season.

Very recently, Wang *et al.*(2004) has pointed out an important role of the existence of strong vertical easterly wind shear over south Asia to the south of the Tibetan Plateau in the northward migration of convective zone with intraseasonal time-scales (including the monsoon onset phase) from the equatorial Indian Ocean toward the Himalayas. This implies 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 an essential element of the Asian summer monsoon system.

3.2 Atmospheric Heating over Tibetan Plateau

The diabatic heating over and around the Tibetan Plateau is attributed to sensible heating from the land surface and latent heating through cumulus convection. Some diagnostic studies based on the MONEX (Monsoon Experiment) date for 1979 (Yanai *et al.* 1992, Li and Yanai 1996, Wu and Zhang 1998) estimated that sensible heating is dominated over the dry land surface of the Tibetan Plateau in the pre-monsoon season, while latent heating through cumulus convection becomes dominant in the wet monsoon season particularly in the eastern part of the Plateau. Ueda *et al.* (2003), based on the GAME (GEWEX Asian Monsoon Experiment) reanalysis data for 1998, re-calculated the diabatic heating rate over and around the Plateau and noticed that latent heating is relatively large even over the western Tibet in the pre-monsoon season, which suggest that latent heating by dry shallow convection along the Himalayan slope and over the Plateau in the pre-monsoon season may also be important for seasonal heating up in the Asian monsoon system. Based on GAME-Tibet intensive observation data for 1998 and a numerical simulation, Kuwagata *et al.* (2001) showed that local mountain-valley circulation systems with diurnal cycles are dominated over the whole of the Plateau during the monsoon season as shown in Fig. 3, and plays an important role in diabatic heating processes over there.

4. Role of Eurasian Snowcover in All-India Monsoon Rainfall (AIMR) Anomaly

Blanford (1884) first noted the relationship between the Himalayan winter snowcover anomaly and the succeeding All-India Monsoon rainfall (AIMR). Walker (1910) followed up his study and re-confirmed this negative correlation. Since 1960s the satellite-based snowcover extent data became available, and so many studies have addressed this issue as a typical example of the role of LAI on the interannual variability of Asian monsoon. Hahn and Shuka (1976) showed an apparent negative correlation between the satellite-based winter snowcover extent anomaly over Eurasia and the following ISMR, though the data used were only 9 years. Since then numerous similar studies have been presented using some different indices of snowcover 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, 2003; Robock *et al.*, 2003). Some studies also discussed Eurasian snow-AIMR relationship, as part of ENSO-monsoon connections (Yasunari, 1987; Kahndekar, 1991; Yasunari and Seki, 1992; Yang, 1996; Yang and Lau, 1998; Kawamura, 1998; Ye and Bao, 2001; Liu and Yanai, 2002).

A central issue for the snowcover-IAMR connection may be how the climate memory of snowcover in winter (or spring) anomaly can be conveyed to the following summer. Some GCM studies attempted to this issue (Barnett *et al.*, 1989; Yamazaki, 1989; Yasunari *et al.*, 1991; Vernekar *et al..*, 1995; Douville *et al.*, 1996; Shen *et al.*, 1998). Though some experiments have given unrealistically large anomalous snowcover (snowmass) as an initial condition, both the albedo and the snow-hydrological effect of anomalous snowcover could play, to some extent, to produce monsoon anomalies in summer. The recent precise observational studies on seasonal march of temperature and circulation field over Eurasia from spring to summer have shown that the influence of snowcover and related soil moisture anomaly on the temperature and circulation anomalies in the lower troposphere is limited primarily when and where snowcover exists seasonally as schematically shown in Fig. 4 (Shinoda *et al.*, 2001; Ueda *et al.*, 2003; Robock *et al.*, 2003). Some studies have concluded, therefore, that the snowcover-AIMR correlation may be an artifact resulted from a common atmospheric circulation pattern which is responsible both for the snowcover anomaly and the AIMR anomaly though the physical process for time-lag of one to a few months between these two parameters has not yet been explained.

5. Role of LAI in the Seasonal Cycle of Australian Summer Monsoon

Australia is a round-shaped continent without complicated topography. It is located in the subtropics and is mostly isolated from other continents. These characteristics make the Australian continent an ideal experimental field in understanding the basics of monsoon dynamics. In this subsection, we focus specifically on the contribution of land-atmosphere interactions to the seasonal transition of the Australian monsoon circulation.

5.1 Differences between Wet and Dry Seasons

Figure 5a shows the difference in climatological monthly precipitation between its maximum and minimum values in annual cycle. Areas where the wet-to-dry difference exceeds 300 mm concentrate in South Asia, especially the southern periphery of the Tibetan Plateau, the northeastern part and west coast of the Indian subcontinent, and the Indochina peninsula. In contrast, only a part of northern Australia satisfies the 300 mm difference in precipitation. As for the Australian summer monsoon, much rainfall is observed over the Arafura Sea, Coral Sea and the northern coast of the Australian continent (not shown), but monsoon rainfall does not penetrate deep into inland, as compared to the South Asian summer monsoon.

In a similar way, we present, in Fig. 5b, the difference in climatological normalized difference vegetation index (NDVI) between its maximum and minimum values in annual cycle. Looking at monsoon regions, the NDVI difference is large in the Indian subcontinent, the Indochina peninsula and southern China. However, northern Australia has no large difference in NDVI. This suggests that the seasonal change of vegetation activity in northern Australia is smaller than that in South Asia.

As suggested from these features, the South Asian summer monsoon system may be more sensitive to land-surface hydrologic processes because its monsoon rainfall intrudes deep into inland. Since the land area of the Australian summer monsoon system is very narrow, on the contrary, its monsoon system might be sensitive to oceanic conditions (such as SST) rather than land-surface conditions. So in that sense, few studies have been done about the impact of the land-surface conditions, such as soil moisture variability, on monsoon variability over Australia. Recently, Timbal *et al.* (2002) explored the importance of soil moisture variability in clarifying the lagged relationship between ENSO and Australian rainfall, and demonstrated that soil wetness fluctuations contribute to an increase in the persistence and the variability of surface temperature and precipitation over the continent. They further noted that a representation of soil moisture variability is a prerequisite for the skill of seasonal forecasting of the Australian climate. Their findings suggest that we cannot disregard the land-surface hydrologic conditions of the Australian continent that also influence differential heating between land and ocean.

5.2 Continent Forcing and its Association with Abrupt Onset

The presence of a continent is expected to affect low-level atmospheric circulations over land and its surrounding ocean, through enhanced land-ocean thermal contrast, prior to monsoon onset. Hung and Yanai (2004) emphasized that such thermal contrast due to differential heating between land and ocean serves as a seasonal preconditioning for the onset. Figure 6 shows the latitude-height composites of the meridional circulation averaged over 115°-150°E before and after the onset day

(Hung and Yanai 2004). Prior to the onset, it can be seen that sensible heating induces a thermally-induced meridional-vertical circulation, characterized by the outflow above 700 hPa and low-level inflow below 925 hPa over the continent. This vertical circulation, which is clearly separated from the ITCZ near the equator, helps to transport low-level moist air inland. A narrow region of descending motion near 15°S is also evident before the onset. Once the monsoon commences, the vertical circulation system induced by the sensible heating and the ITCZ system merge together, and the region of descending motion is replaced by that of ascending motion.

The premonsoon circulation features that they pointed out are also noted by Kawamura *et al.* (2002). During the premonsoon period, the sensible heating from the land surface results in a continental-scale thermal low at the lower level below 850 hPa and a thermal high at 600-700 hPa level, in conjunction with the shallow vertical circulation (see Fig. 6). The dominance of these circulations leads to dry intrusion into the layer at ~700 hPa over the Arafura Sea and Coral Sea through the horizontal and vertical advective processes. Since SSTs rapidly increase in the Arafura Sea and Coral Sea prior to the onset, a combination of the SST increase and the dry intrusion creates a more potentially unstable condition. While the subsidence in the periphery of the continent suppresses convection, potentially convective instability is further intensified. If large-scale disturbances with ascending motion, such as the Madden-Julian oscillation (MJO), arrive at the domain where the lower troposphere is potentially unstable, deep cumulus convection breaks out suddenly, implying the onset of the monsoon. The role of MJO in the onset process is also documented by Hendon and Liebmann (1990) and Hung and Yanai (2004).

Continent forcing (sensible heating) is critical in the onset mechanism proposed by Kawamura *et al.* The continent forcing is also able to change surface wind stress on the ocean surface in the periphery of the continent, through the effects of geostrophic balance and surface friction (e.g., Xie and Saiki, 1999), resulting in anomalous SST and surface ocean currents. In the case that a continent is away from the equator, prior to the onset of the summer monsoon, westerly wind stress anomalies are induced due to geostrophic balance over the ocean off the equatorial side of the continent where trade winds dominate. Attenuated trade winds lead to decreased surface evaporation and thus increased SST. As a result, very high SSTs, different from other oceanic regions, cover the ocean off its equatorial side before the onset, which also acts as an important preconditioning for the abrupt onset (see Fig. 6 again). To fully understand why monsoon onset is abrupt, thus, we need to consider not only the land-atmosphere coupling over the continent but also the ocean-atmosphere coupling in the vicinity of the continent. Minoura *et al.* (2003) demonstrated that Kawamura *et al.*'s onset mechanism could apply to the abrupt onset of the Indian summer monsoon at the beginning of June.

As has been discussed in section 4, the land-surface hydrologic conditions of the Asian continent prior to the monsoon onset are likely to affect the following Asian monsoon variability. We need to examine whether or not such land-surface conditions during premonsoon period are also important in the Australian summer monsoon variability. Likewise, soil wetness is expected to play an influential role in the persistence of monsoon rainfall after the onset through a feedback process. Further studies are required on the significant impact of soil moisture variability on the Australian summer monsoon system in terms of various timescales (e.g., intraseasonal and interannual timescales).

6. New Aspects of LAI in the Monsoon System

6.1 Role of Vegetation

As has been discussed in section 2, the time-scale of climatic memory effect of soil moisture is likely to be around 2-3 months (or less). However, this time scale may be changed to further longer, if we include the role of vegetation with deep root zone and control mechanism of evapo-transpiration. The long-term observations of surface energy and water fluxes in some vegetated regions conducted as part of GAME-AAN (automatic AWS Network) have revealed some new roles of vegetation in climate memory effect. In the evergreen monsoon forest in northern Thailand, for example, the seasonal maximum of evapo-transpiration appeared in the midst of the pre-monsoon dry season (March to April) which suggest that soil moisture fed by monsoon rainfall in the previous year is maintained for more than a half year in a deeper zone (possibly several meters depth or more) and is gradually transpirated from the forest leaves depending upon the atmospheric condition (Tanaka *et al.*, 2003; 2004). On the other hand, in the Siberian cold region in Taiga, the seasonal cycle of energy and water fluxes from the surface was completely controlled by phenology of Taiga forest combined with melting and refreezing of permafrost; latent heat flux started at the foliation and ended at the falling of leaves (Ohta *et al.* 2001). These results suggest that continental-scale vegetation may be in reality a key parameter for controlling surface energy processes. Some recent GCM and regional model studies (e.g., Kanae *et al.* 2003, Shinoda and Uyeda 2002, Xue *et al.* 2004, Wang *et al.* 2001, Yasunari *et al.* 2004) also have suggested that even the regional and continental-scale monsoon circulation and precipitation change considerably by changing the vegetation parameters (albedo, field capacity, evapotranspiration efficiency etc.)

6.2 Role of Frozen Ground and Permafrost

Another new process to be included in LAI is a role of permafrost and melting/ freezing process of surface and subsurface layer in controlling seasonal energy and water cycle. As described in the previous subsection, Taiga and permafrost is likely to be a combined system (Ohata *et al.* 2001). Recently, water cycling processes of this system has been proved using the isotope hydrological method including precipitation, evapotranspiration from the Taiga forest, water transport through root zone and water movement in the seasonally-melting zone of permafrost, which suggests the soil moisture anomaly at near-surface layer has a memory effect of one year or longer (Sugimoto *et al.* 2003). Takata and Kimoto (2000) conducted a GCM simulation of seasonal cycle of the Eurasian continent, using a land surface model including a more realistic soil-freezing/melting process with impermeability of water in the frozen layer. They found that by including these processes surface temperature increased considerably in summer due to decrease of soil moisture and suppression of evaporation, which in turn strengthen the monsoon circulation in the tropics and subtropics. The cause of the soil moisture decreasing in summer is resulted from increase of runoff in spring due to the existence of impermeable layer of water in the near sub-surface layer. These processes are thought to be more realistic which most of the land surface models have not treated yet. These processes may be essential not only in Siberia but also on the Tibetan Plateau. Xu *et al.* (2001) estimated surface energy and water balance on the Tibetan Plateau using many station-based data, and found the efficiently large runoff and sensible heat (e.g., suppression of latent heat) from dry surface in the pre-monsoon season. Under this climate condition, freezing/melting process with its seasonal timing may effectively control the surface Bowen ratio, which in turn should be important for interannual variability of atmospheric heating over there. The recent long-term observation of surface energy balance on the center of the Plateau for several years as part of GAME-Tibet (CAMP-Tibet) has supported the importance of these processes (Ishikawa *et al.* 2004).

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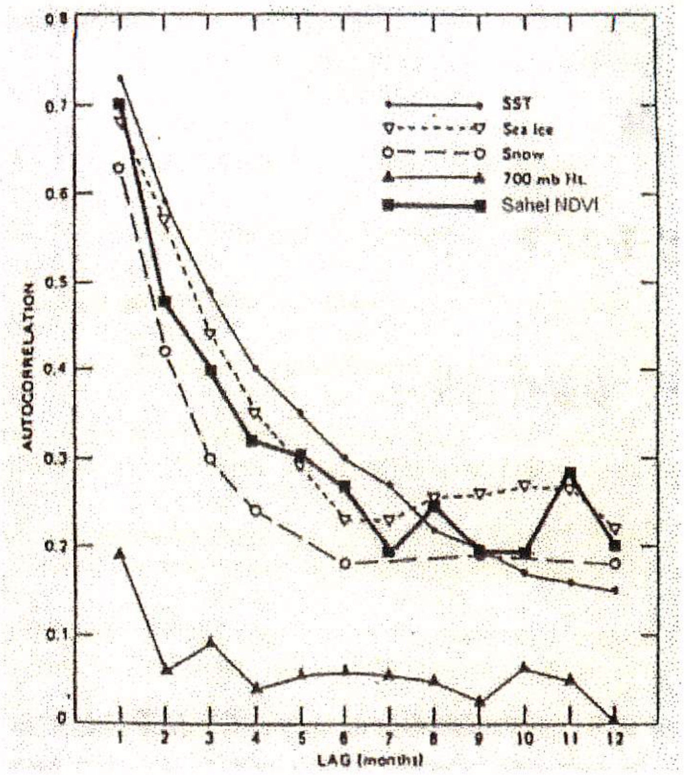

Figure 1. Autocorrelation of monthly anomalies for northern Pacific SST, arctic sea ice, Eurasian snowcover and 700hPa height (Walsh *et al.* 1985). That for NDVI anomalies at Sahel region, Africa (Shinoda and Gamo 2000) is also plotted.

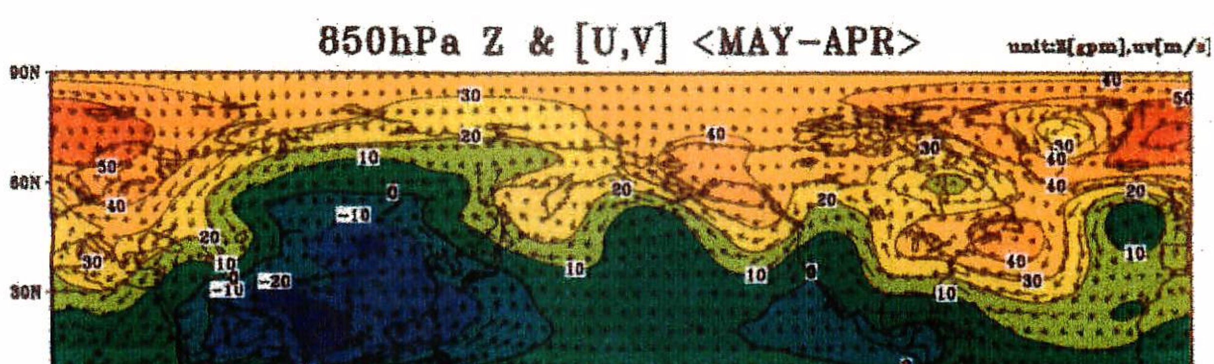

Figure 2. Difference of monthly mean 850hPa height climatology between April and May. Negative (positive) areas show height decreases (increases) from April to May.

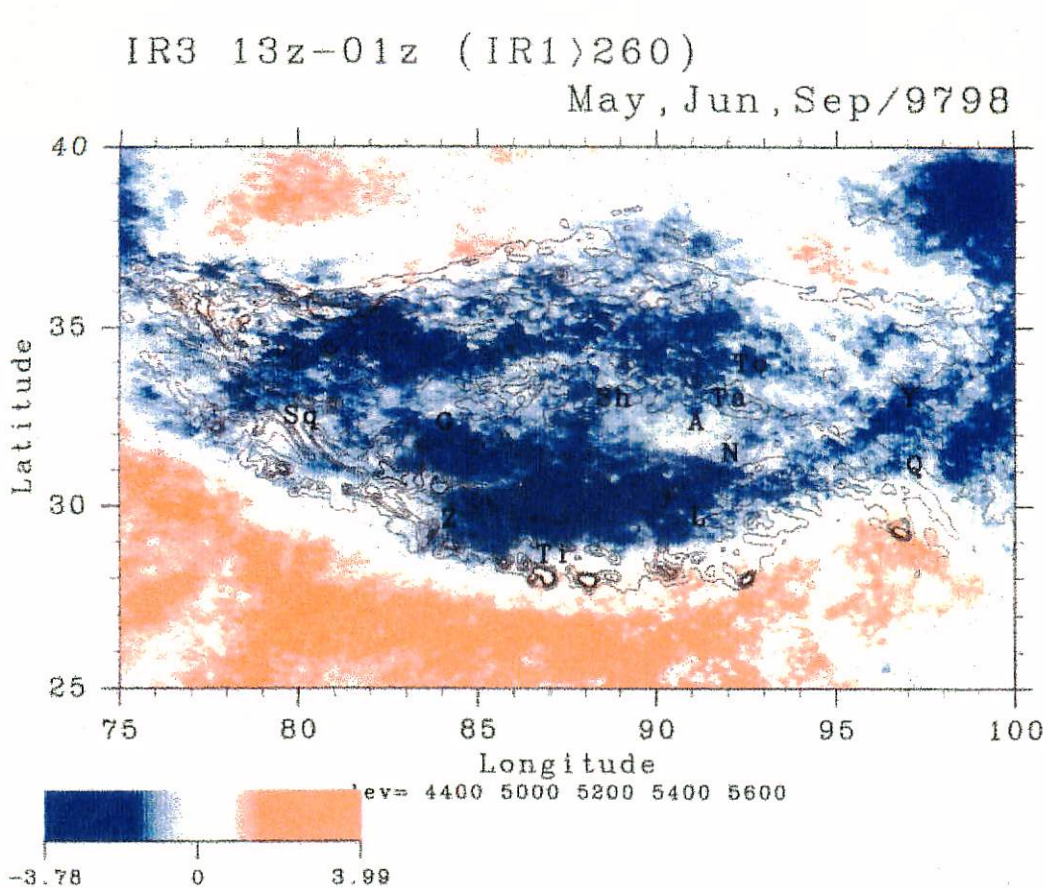

Figure 3. Difference of brightness temperature of water vapor channel of GMS-5 satellite between 1300 UTC and 0100 UTC (1300 UTC – 0100 UTC) averaged for May, June and September of 1997 and 1998. . Only the pixels greater than 260K are included. Contours show altitudes of 4400, 5000, 5200, 5400 and 5600m MSL. (Kuwagata *et al.* 2001).

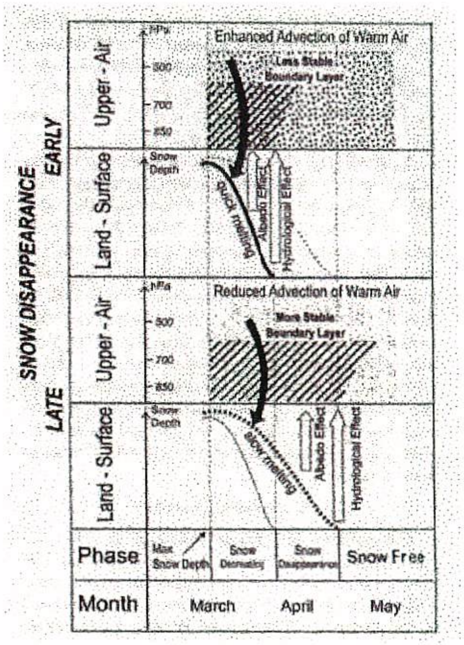

Figure 4. Schematic diagram illustrating snowcover and atmosphere interactions over central Eurasia for the early and late snow disappearance. Hatching (dotting) indicates vertical and seasonal ranges of stable boundary layer (advection of southern warm air towards central Eurasia). Thick black (white) arrows denote the effect of atmosphere on snow cover (the effect in the reversed direction). (Shinoda *et al.* 2001).

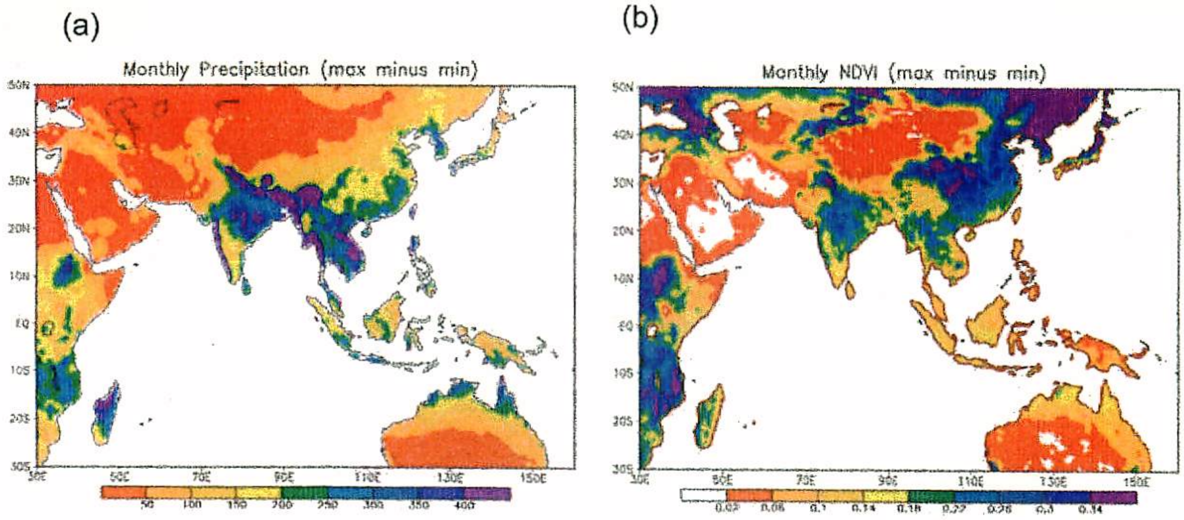

Figure 5. (a) Difference in climatological monthly precipitation between its maximum and minimum values in annual cycle. The data used are the monthly total precipitation series (Version 1.02) produced by Cort J. Willmott and Kenji Matsuura. Unit is mm. (b) As in (a) but for normalized difference vegetation index (NDVI).

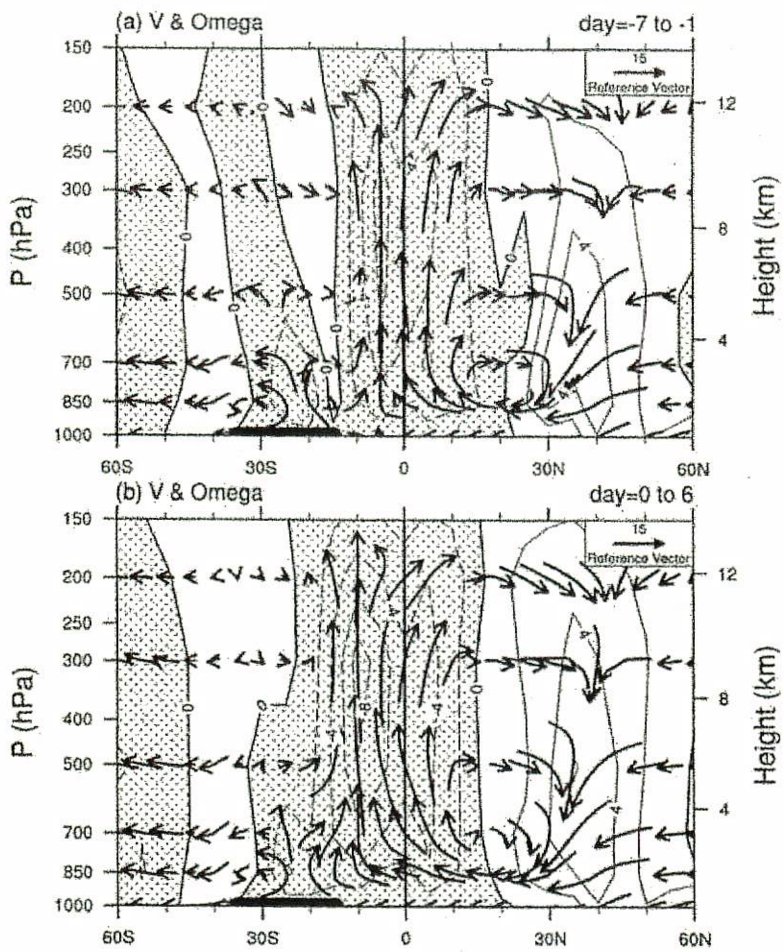

Figure 6. 1979-93 latitude-height composites of the meridional circulation (v, omega; arrows) and values of omega (Pa s^{-1}) averaged over 115°-150°E: (a) from 7 days to 1 day before the onset, and (b) from the onset day to 6 days later. Areas of upward motion (negative omega) are shaded. (Hung and Yanai 2004.)

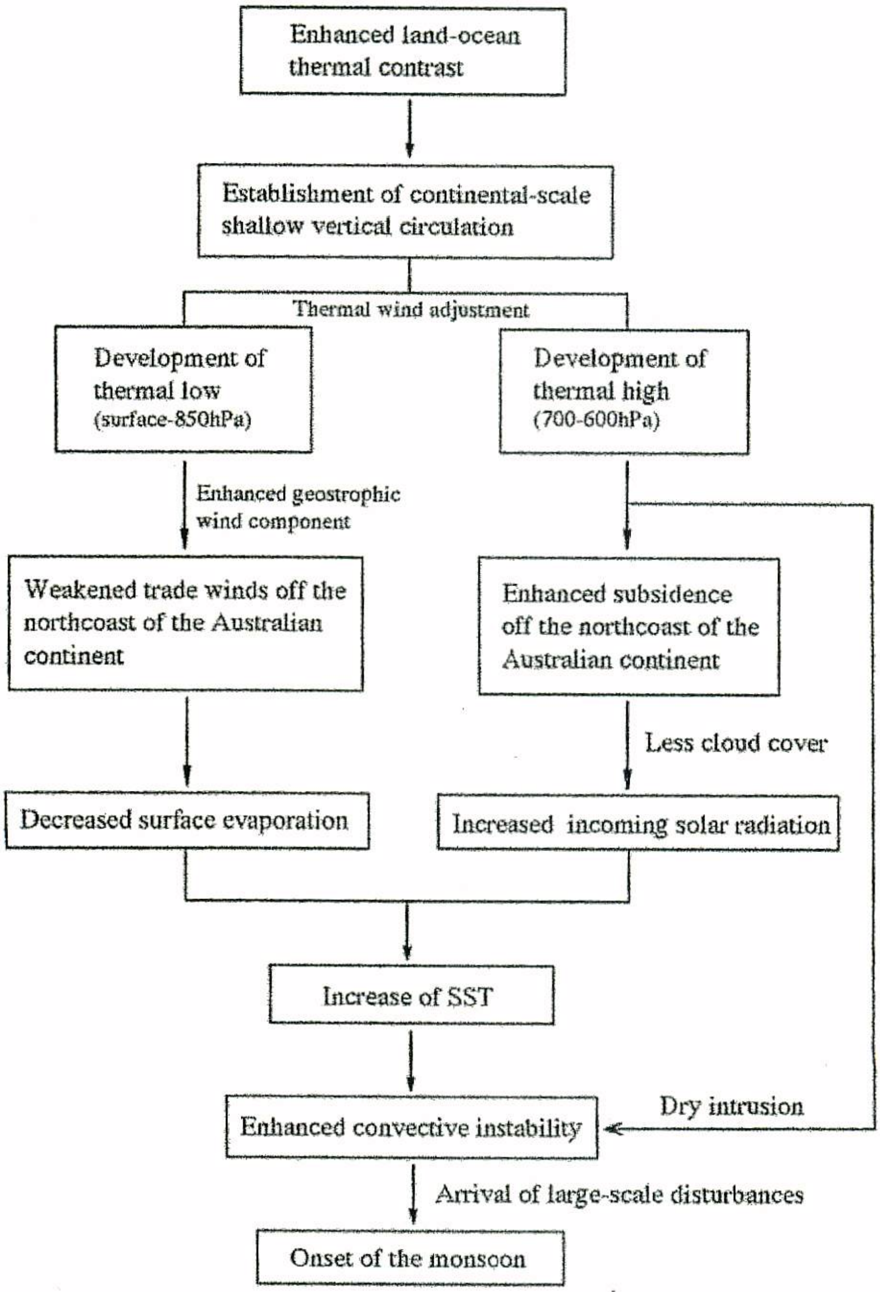

Figure 7. Schematic diagram representing what dynamic and thermodynamic processes are crucially responsible for the occurrence of the onset of the Australian summer monsoon. (Kawamura *et al.* 2002)