Entropy Evolution Characteristics Associated with the Development of the South Asian Monsoon

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ABSTRACT

The structure and evolution characteristics of atmospheric entropy production associated with the climatologic monsoon onset and evolution were investigated using the National Centers for Environmental Prediction (NCEP) reanalysis data. The entropy balance equation contains two parts. The first part is internal entropy production that corresponds to natural dissipation. The second part is external entropy production that is associated with lower-boundary entropy supply. It is shown that the dissipation process represented by internal entropy production can be used to describe the thermal and dynamical structures of the monsoon. The thermal dissipation due to turbulent vertical diffusion and convection is highly correlated to precipitation. The dynamic dissipation due to wind stress becomes very strong over the Arabian Sea and southwestern part of India in boreal summer, and dynamic dissipation can describe the monsoon structure more clearly than variables such as wind shear. The correlation between surface entropy supply and internal entropy production is so large that the surface entropy supply can also be used to evaluate the monsoon. Over the desert region of Rajasthan, the dissipation is relatively weaker than its surroundings owing to descending large-scale eddy flow and a weak convective flux. The analysis of atmospheric entropy provides a new way to describe the monsoon development characteristics, which differs from those derived from a traditional analysis method.

1. Introduction

Nonequilibrium thermodynamics using entropy as its main tool has been widely used in many scientific fields. The entropy balance equation contains two parts: internal entropy production, which corresponds to dissipation, and external entropy production, which corresponds to boundary entropy supply. In a given climate system, dissipation to orderliness occurs spontaneously. This is an irreversible process accompanied by positive internal entropy production. It is the environment that provides the system with a negative entropy supply to prevent the system from reaching a state of maximum entropy characterized by nonorderliness. Observations show that the atmosphere is an orderly system with well-organized thermal and dynamical structures. Therefore, atmospheric entropy must play a certain role in any climate or weather event. In the last several decades, the entropy related to the second law of thermodynamics has also been applied to atmospheric science (e.g., Li et al. 1994; Pauluis and Held 2002; Ozawa et al. 2003; Pascale et al. 2009; Lucarini et al. 2011). Recently, the atmospheric entropy balance and the climate dissipation structure have been investigated in Li and Chylek (2012). In Li et al. (2014), the dissipation structure of extratropical cyclone is explored. In this work, we will focus on the application of atmospheric nonequilibrium thermodynamics to the Asian monsoon.

The term “monsoon” was originally defined as a reversal of prevailing wind direction between winter and summer (Ramage 1971), and later it was often referred as a contrast of precipitation between a wet and dry season (e.g., Tao and Chen 1987; Wang and LinHo 2002).
As the most energetic system on Earth, the Asian monsoon exerts a great impact on global climate (Das 1972; Rao 1976; Chang and Krishnamurti 1987; Ding 1994; Wang 2006). The Asian monsoon system consists of three subcomponents: South Asian monsoon (SAM), East Asian monsoon (EAM), and western North Pacific monsoon (WNPM). SAM and EAM are typical continental monsoons driven by land–ocean thermal contrast. WNPM is an oceanic monsoon driven by hemispheric asymmetric SST gradient.

The time evolution of the Asian monsoon has distinctive characteristics. Large-scale convection develops first over the southern Bay of Bengal in late April and then propagates northward (Li et al. 2013). Both the SAM and EAM begin to withdraw southward in August to September (Wang and Lin Ho 2002). WNPM is characterized by eastward propagation of convective rain from mid-May to August (Wu and Wang 2001). The details for the time evolution of the Asian monsoon are shown in a review paper by Li (2013).

Besides a pronounced annual cycle, the Asian monsoon also exerts remarkable intraseasonal variation (e.g., Yasunari 1979; Krishnamurti and Subrahmanyan 1982) and interannual variation (e.g., Walker 1923). For example, during the boreal summer, there is clear northward propagation of intraseasonal convective rainband over the northern Indian Ocean and South China Sea (Jiang et al. 2004). In addition, pronounced westward propagation of intraseasonal rainbands is observed over the off-equatorial western Pacific (Li and Wang 2005). The El Niño–Southern Oscillation (ENSO) has been considered as a major factor affecting the monsoon interannual variability. Walker (1923) first recognized the effect of the Southern Oscillation on SAM. Since then, a number of studies have been conducted to elucidate the monsoon–ENSO relationship (e.g., Webster and Yang 1992; Ju and Slingo 1995). SAM tends to have a simultaneous negative correlation with the eastern Pacific SST (e.g., Rasmusson and Carpenter 1983; Wang 2006). The Asian monsoon system consists of three subcomponents: South Asian monsoon (SAM), East Asian monsoon (EAM), and western North Pacific monsoon (WNPM). SAM and EAM are typical continental monsoons driven by land–ocean thermal contrast. WNPM is an oceanic monsoon driven by hemispheric asymmetric SST gradient.

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The change of atmospheric heat sources is a key factor in the abrupt transition from a dry season to a rainy season (Krishnamurti 1985; Li and Yanai 1996; Chou et al. 2003; Chen 2006). The onset of the summer monsoon brings with it a strong thermal and dynamical structure. According to the dissipation theory by Prigogine (1969), internal entropy production (or dissipation) process must exist for the orderly thermal and dynamical structure of the monsoon. At the same time, the negative entropy sink must be engaged in maintaining the orderly structure of the monsoon by balancing the positive entropy production in the dissipation process. Therefore, the examination of internal entropy production and evolution can help understand the development of a monsoon system.

The purpose of this study is to understand the characteristics of the Asian monsoon based on nonequilibrium thermodynamics using atmospheric entropy theory as a tool. In traditional Asian monsoon study, the attention was mostly paid to the seasonal reversal change in precipitation and wind, which are the main characteristics of Asian monsoon. Nonequilibrium thermodynamics will address the dissipation process to orderliness of a system as indicated by internal entropy production. In the following, we will show that the main characteristics of Asian monsoon can also be clearly explored by study of monsoon dissipation structures. Although ENSO is known to have a pronounced effect on the strength of SAM, we do not discuss such a relationship in this study. Analyzing ENSO from nonequilibrium thermodynamics is beyond the scope of this work since ENSO is associated with ocean circulation. Investigating the ENSO role on SAM will be a subsequent work. Though the Asian monsoon extends to East Asia and the western North Pacific, we will mainly focus on the South Asian monsoon.

2. Theoretical background

Atmospheric entropy can be represented by (Emanuel 1994)

$$s = c_{pd} \ln \theta + \frac{\lambda q_v}{T} - q_v R_v \ln H,$$  \hspace{1cm} (1)

where $s$ is the entropy per unit mass, $\theta$ is the potential temperature, $T$ is the temperature, $c_{pd}$ is the specific heat at constant pressure, $R_v$ is the gas constant of water vapor, $\lambda$ is the latent heat of evaporation of water, and $H$ is the relative humidity. Based on (1), Li and Chylek (2012) obtained the entropy balance equation

$$\frac{\partial (\rho s)}{\partial t} = - \frac{\partial}{\partial x_i} \left( J_{\theta i} + J_{H i} - R J_{H \theta} \ln H \right)$$

$$- \frac{\partial}{\partial z} \left( F_h T + F_l T + F_c T - R V_{T} + F_{cm} \ln H - \rho u \tau_{\alpha} \right)$$

$$+ \alpha + c_{pd} \rho \frac{\partial u}{\partial t},$$  \hspace{1cm} (2)

where $\frac{\partial}{\partial t} = \frac{\partial}{\partial t} + \nabla_i (\partial / \partial x_i)$ is the Lagrangian derivative; $\rho$ is air density, $J_{\theta i} = c_{pd} \rho \theta \overline{u_i}$ and $J_{H i} = \lambda q_v \overline{u_i}$ are large-scale-eddy thermal flow and moistening heat flow,
respectively; \( q_v \) is the specific humidity for water vapor; \( F_h, F_l, \) and \( F_c \) are the small-scale mean sensible heat flux, latent heat flux, and convective heat flux, respectively; \( F_{cm} \) is the convective moisture flux corresponding to the moistening heating in convection; \( \tau_{ij} \) is the mean diffusion acceleration tensor, and \( \bar{Q}_r \) is the radiative heating rate. The subscript \( i \) (or \( j \)) takes values 1, 2, and 3 representing two horizontal directions and one vertical direction. For example, \( x_1, x_2, x_3 = x, y, z \) for coordinates and \( v_1, v_2, v_3 = u, v, w \) for velocities in the \( x, y, z \) directions. The subscript \( \alpha \) takes the values 1 and 2 representing the two horizontal directions. The convention of summing the repeated index \( i \) (or \( j \)) from 1 to 3 and the repeated index \( \alpha \) from 1 to 2 is adopted. A bar over a physical variable represents the time mean. In (2), \( \sigma \) is internal entropy production,

\[
\sigma = \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{L} + \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{R} = \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{F}_h + \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{F}_l + \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{F}_c + \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{R}_w - \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{R}_l - \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{F}_{cm} + \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{R}_w \ln \bar{H} + \frac{1}{\rho} \frac{\partial}{\partial \alpha} \bar{F}_{cm} \ln \bar{H},
\]

where \( \bar{C} \) and \( \bar{E} \) are the condensation and evaporation rates, respectively. For an individual time step, balance equation is as in (2), but without considering the large-scale eddy. The effects of a large-scale eddy coming from the temporal variation only appears in a climate study with a considerably longer time scale but does not apply to the events with short time scales.

Equation (2) shows that the time evolution of entropy is caused by the divergence of external entropy flux, the internal entropy production, and an entropy sink due to radiation. This follows since generally \( c_{pd} \bar{Q}_r / T < 0 \) in the atmosphere. In Li and Chylek (2012), it is shown that the radiative entropy sink plays an important role on the global entropy balance in the atmosphere.

### 3. Dissipation structure of the South Asian monsoon

Internal entropy production illustrates the relationship between the gradient of mean fields (temperature, potential temperature, and moisture as shown in the logarithm of relative humidity) and the resolved large-scale eddy flows and unresolved small-scale diabatic heat fluxes. Physically, internal entropy production represents the dissipation process (Prigogine 1969; deGroot and Mazur 1984; Hu 2002). As shown in (3), the typical form of internal entropy production is

\[
\sigma = F \frac{\partial}{\partial \alpha} \frac{1}{T},
\]

where \( F \) is the thermal flow. If the thermal flow follows the diffusion law (Fourier’s law) (deGroot and Mazur 1984), then

\[
F = -k \frac{\partial T}{\partial \alpha},
\]

where \( k \) is the diffusivity, and hence

\[
\sigma = k \frac{\partial T}{T^2} \frac{\partial T}{\partial \alpha} > 0.
\]

Equation (5) shows that thermal flow moves along the gradient of temperature from a higher-temperature region to a lower-temperature region. Thus, the flow tends to decrease the gradient of the temperature field by extracting energy from the high-temperature regions and depositing it into the low-temperature regions. This is an irreversible process that results in a reduction of the temperature gradient. Therefore, the positive internal entropy production is referred to as dissipation (Prigogine 1969).

In the atmosphere, latent heat flux and convective heat flux do not follow Fourier’s law in (5).

However, if a thermal flow is upward, thereby downward to the vertical temperature, then the dissipation process discussed above is held true and positive internal entropy production is given by

\[
\sigma = -k \frac{F}{T^2} \frac{\partial T}{\partial \alpha} > 0,
\]

since in the troposphere \( \partial T / \partial \alpha < 0 \).

In the atmosphere, structure in the temperature field is always present because of the nonhomogeneous distribution of incoming solar heating. Regionally, the air–sea interactions at the western boundary currents also help maintain the atmospheric temperature structure owing to strong ocean fronts there. In addition, the land–sea heating contrast also plays a role. A region of positive internal entropy production must correlate with the dissipative temperature structure. Otherwise, the dissipation process cannot last long. Dissipation is a common process in nature. Dissipation is due not only to the temperature structure but also to the dynamic and moisture structures. The details will be addressed in the following.

The monsoon is a gradually developing climate system (Ramage 1971; Das 1972; Chang and Krishnamurti 1987; Wang 2006). The South Asian summer monsoon
usually starts in May and ends in September (Wang and LinHo 2002). During the 5-month period, the monsoon generally has strong thermal and wind structures and is accompanied by heavy precipitation. Therefore, the orderliness in temperature, moisture, and dynamical structures over South Asia changes in different monsoon stages. As discussed above, dissipation always occurs to the orderly structure. Therefore, the study of dissipation process can be indicative of the development stage for a monsoon.

We use 10-yr National Centers for Environmental Prediction (NCEP) monthly-mean data from 1998 to 2007 to evaluate the climate process of monsoon. The monthly data were interpolated onto a $192 \times 94$ Gaussian grid with 18 vertical levels. In isobaric coordinates, pressure at a specific model level is $p = p_s \eta$, where $p_s$ is the surface pressure and $\eta$ is the ratio of a specific model level to the lowest model level.

The left column of Fig. 1 shows the distribution of vertically integrated internal entropy production based on (3), with the contribution from both large-scale eddy flows and small-scale diabatic heat fluxes. The vertical integration ranges from $\eta = 1$ to $\eta = 0.01$, where $\eta = 0.01$ is the top level of NCEP data. The monthly-mean patterns for January and for May–September are shown. Compared to the summer monsoon, the winter monsoon is much weaker and mostly occurs in December and January. Here, we use January to represent the winter monsoon period.

In the premonsoon month of May, internal entropy production appears over the Arabian Sea, Bay of Bengal, and Indian subcontinent (Fig. 1). The dissipation structure starts to establish. In the northwestern Pacific, dissipation is relatively weak. However, large dissipation contours have begun to appear in the South China Sea. The moist air over the Pacific moves northward and meets the cooler continental air mass (Ding 1994). This forms frontal depressions and brings precipitation to southeastern China (called mei-yu season). The moist airflow moves northward toward China creates a downgradient of temperature, which causes an increase of internal entropy production.

In June, internal entropy production is more enhanced over the Arabian Sea and the Bay of Bengal. This is where the two main monsoon branches are located. Additionally, internal entropy production is relatively large over the west coast of India and, in particular, the mountainous Ghats region. Therefore, dissipation is strongly associated with the climate process of the monsoon. In the tropical Indian Ocean, large internal entropy production also appears. This is caused by the dissipation from strong convective flux (see details in the middle column of Fig. 1).

In July, the contour center of internal entropy production in the Arabian Sea has reached southwestern India. At the same time, dissipation over the Bay of Bengal has been strengthened and has extended into southeastern India. Over the desert region of Rajasthan, the dissipation is relatively weak compared to its surrounding areas. Detailed analysis shows that weak dissipation is attributed to the downward large-scale eddy flows and weaker upward convective heat flux (see discussion for Fig. 7).

In August, internal entropy production starts to withdraw from India over the Arabian Sea and the Bay of Bengal. Especially in the Arabian Sea, the signal of internal entropy production becomes much weaker compared to that of July.

In September, the values of internal entropy production over the Arabian Sea and the Bay of Bengal become even weaker. Additionally, the big contour over the equator in the Indian Ocean retreats to the Southern Hemisphere. The summer monsoon is close to an end.

To understand the detailed physics of dissipation, we consider contributions from different physical processes. The contribution of internal entropy production due to turbulent vertical diffusion and convection is

$$\sigma_t = \langle \tilde{F}_h + \tilde{F}_l + \tilde{F}_c \rangle \frac{\partial}{\partial T} \frac{1}{T}.$$

We refer to this part of internal entropy production as thermal dissipation. Thermal dissipation displays the physics that the sensible, latent, and convective fluxes move downgradient to temperature. This process dissipates the temperature gradient.

In the middle column of Fig. 1, the vertically integrated contributions of thermal dissipation are shown for the same 6 months as the left column of Fig. 1. From May to September the changes of contours clearly show the onset and withdrawal of summer monsoon over the Arabian Sea, the Bay of Bengal, and the Indian subcontinent. The main contour patterns are similar to that of the total internal entropy production shown in the left column. This indicates that the thermal part dominates the internal entropy production. The contributions from the large-scale eddy and other processes are relatively small. For the winter monsoon, it is found that thermal dissipation is weak over the Indian subcontinent, similar to that of the total internal entropy production. This is because the dry condition produces weaker convective flux.

In the right column of Fig. 1, the distribution of precipitation is shown for the same period. It is found that the distributions of thermal dissipation and precipitation are similar. A factor of the monsoon is that the
airstream is laden with moisture. In the thermal dissipation process, the upward thermal flow can drive air rising to reach saturation, which causes precipitation. Therefore, thermal dissipation is associated with precipitation. From June to August, there is heavy precipitation on the western coast of India (Ramage 1971; Lau and Wu 2001). This is caused almost entirely by the topographic barrier presented by the shape of the Western Ghats. This leads to strong upward thermal flow and large precipitation.

The strong seasonal changes of precipitation and wind are two distinguishing features of the South Asian monsoon. We show subsequently that the feature of precipitation can be properly described by thermal dissipation.

**FIG. 1.** (left) Distribution of vertically integrated internal entropy production (0.01 W m$^{-2}$ K$^{-1}$) for 6 months. (middle) As in the left column, but for thermal dissipation. (right) Distribution of precipitation ($10^{-5}$ kg m$^{-2}$ s$^{-1}$).
Besides the climatological-mean result, Fig. 2 shows the daily evolution of thermal dissipation for the period 3–7 August 2005. The reason for choosing these specific 5 days is that a typhoon in the western Pacific was developing during this period. For an individual time step, the contribution of the large-scale eddy is not considered as discussed above. The left column of Fig. 2 shows the distribution of thermal dissipation with a contour interval of 0.1 W m\(^{-2}\) K\(^{-2}\) and the right column shows precipitation with a contour interval of 10\(^{-5}\) kg m\(^{-2}\) s\(^{-1}\) during 3–7 Aug 2005.

**Fig. 2.** Daily evolution maps of (left) thermal dissipation (contour interval: 0.1 W m\(^{-2}\) K\(^{-2}\)) and (right) precipitation (contour interval: 10\(^{-5}\) kg m\(^{-2}\) s\(^{-1}\)) during 3–7 Aug 2005.
vertical integration from $\eta = 1$ to $\eta = 0.01$. In the right column of Fig. 2, the distribution of precipitation is shown. The daily distribution of precipitation is similar to that of the thermal dissipation. This result is the same as the climatology mean result shown in Fig. 1.

At 0000 UTC 3 August, there was a typhoon near Taiwan. The typhoon was a tropical cyclone moving northwestward, mainly targeting China and Japan. At 5–6 August this typhoon had entered China and its strength had gradually decreased. A typhoon is usually accompanied by heavy rain. From the distribution of precipitation, the development and track of the typhoon toward China was clearly evident. The track of this typhoon can be indicated by thermal dissipation, especially in its early stage from 3 to 5 August. When the typhoon made landfall, the upward thermal flows, especially the convective flux, were much reduced owing to the less moist condition. Therefore, the thermal dissipation became less evident. However, the precipitation signal of the typhoon was still clear after landfall. The precipitation water was carried on by the typhoon on its track.

We can further split thermal dissipation into convective dissipation and latent–sensible dissipation, corresponding to $F_c$ and $F_h + F_l$, respectively. These dissipations are shown in the left and middle columns of Fig. 3. It is seen that the distribution of convective dissipation is consistent with the development and track of the typhoon.
dissipation is very similar to that of thermal dissipation. This means the convective heat flow is much larger than the latent–sensible heat flow (Johnson 2006). In Fig. 3, the plot scale of latent–sensible dissipation is one order smaller than that of the convective dissipation. There is relatively large latent–sensible dissipation over the Arabian Sea and the Bay of Bengal, and this is different than the convective dissipation. The track of the typhoon is also clearly indicated by the convective and latent–sensible dissipation, especially in the early stage of 3–5 August. It is found that the latent–sensible dissipation is mostly surrounding the typhoon. In contrast, the convective dissipation occurs mostly in the core center of the typhoon. This is very different from the result of an extratropical cyclone, in which the convective dissipation occurs mostly in the leading edge (Li et al. 2014).

Dissipation occurs not only to the temperature structure (i.e., temperature gradient) but also to the dynamical and moisture structures. In Fig. 4, the left column displays the contribution of the vertically integrated internal entropy production due to the wind stress; that is,

$$\sigma_d = \rho u^2 \frac{\partial \tau_{3,a}}{\partial z}.$$

We term the internal entropy production by wind stress as dynamic dissipation. In a dynamic dissipation process, temperature plays a weak role since in the lower atmosphere the vertical variation of $\tau_{3,a}$ is much larger than that of temperature. Therefore,

$$\sigma_d \approx \rho \frac{\bar{u}_1}{T} \frac{\partial \tau_{1,3}}{\partial z} + \rho \frac{\bar{u}_2}{T} \frac{\partial \tau_{2,3}}{\partial z}.$$

Thus, the dynamic dissipation shows a dissipation to dynamic structure of wind stress by horizontal wind of $\tau_a$. Since $\tau_{3,a}$ represents the horizontal momentum in the vertical direction, $-\pi_o \frac{\partial \tau_{3,a}}{\partial z}$ represents the down-gradient of momentum induced by the horizontal wind. This is similar to the physical process shown in (4).

In the middle and right columns of Fig. 4, the climatological-mean wind shear fields in longitudinal and latitudinal directions are presented. The definition of wind shear is the wind speed at 250 hPa minus the wind speed at 850 hPa. Wind strength and direction are the most significant signals of the monsoon. Dynamic dissipation, which contains the factors of wind speed, wind direction, and wind stress, is a comprehensive physical quantity describing the dissipation process associated with wind. The left column of Fig. 4 shows the climatological monthly-mean results for vertically integrated distributions of dynamic dissipation. In May, dynamic dissipation appears in western India but the corresponding result cannot be found in the $u$ component; only the $v$ component of wind shear shows weak signal over western India.

Monsoon wind is most pronounced in the summer season (Das 1972; Chang and Krishnamurti 1987; Wang 2006). In June, very strong dynamic dissipation sweeps across the Arabian Sea. The front part of dynamic dissipation gradually advances northeastward to Bombay. The corresponding results can be found in both the zonal and meridional components of wind shear. This suggests that the summer monsoon wind is toward the northeastern direction, which matches the results of dynamic dissipation. Additionally, the correlation between dynamic dissipation and wind shear can be found in the Bay of Bengal. Over the Indian subcontinent the dynamic dissipation is most visible over the regions of the mountainous Ghats, northern India, and the Himalayan Mountains, but such features are not clearly shown in the wind shear. Therefore, dynamic dissipation is more relevant in describing the strength and variation of monsoon wind over the Indian subcontinent.

In July, the whole Indian subcontinent comes under the grip of the summer monsoon, as clearly shown by dynamic dissipation. In August, the monsoon over the Arabian Sea and Indian subcontinent starts to withdraw as the dynamic dissipation becomes weaker. In September, the dynamic dissipation becomes very weak over the Arabian Sea and the Bay of Bengal. Therefore, dynamic dissipation is more likely to occur in the peak period of the monsoon. In the peak period, the monsoon dynamics can support a large horizontal wind that dissipates the existing dynamic (wind stress) and temperature structures.

For the winter monsoon in January, there are two regions with large dynamic dissipation. One is over the Himalayan Mountains and the other is near the Somali coast. In the winter season, the monsoon wind is north-easterly from northern India and the Himalayas. Thus, the dynamic dissipation is consistent with winter monsoon characteristics. However, this specific feature near the Himalayan region is not shown in the $u$ component of wind shear and only is weakly shown in the $v$ component. The dynamic dissipation is a three-dimensional quantity containing the wind stress, which associates the horizontal motion with the vertical motion. Therefore, the dynamic dissipation produces very different results over the Himalayas compared to that of the wind shear, which only addresses the horizontal components. In the Himalayas the upward air motion is important (Ramage 1971; Wang 2006). Off the Somali coast, there is no signal in the $u$ component of wind shear and only
a very weak signal in the $v$ component. However, the signal of dynamic dissipation is clear over the Somali coast regions.

Figure 5 shows the daily evolution of the monsoon during 3–7 August 2005, as in Figs. 2 and 3. The left column of Fig. 5 shows the distribution of dynamic dissipation, with vertical integration from $\eta = 1$ to $\eta = 0.01$. Note that strong dynamic dissipation is found in the Arabian Sea, from the Somali coast extending to the southwestern Indian subcontinent. Over the Bay of Bengal, the dynamic dissipation is also noticeable. In the middle column of Fig. 5, the distributions of wind shear...
(the zonal component) are shown. It is found that the wind shear is also indicative of the monsoon structure, especially in both the Arabian Sea and the Bay of Bengal (Jiang et al. 2004).

Over the region of the Himalayas, there is strong dynamic dissipation mostly owing to the uprising air mass, but the corresponding result cannot be found from wind shear. Also, in the Turkish–Iranian mountain zone there is strong dynamic dissipation, but with little signal in wind shear.

For the typhoon as discussed in Figs. 2 and 3, it is found that there is no typhoon signal indicated in the wind shear. A typhoon is accompanied by a cyclic-rotation structure and a minimum surface pressure in its center. Using the pressure anomaly to study typhoons is a very common approach. The low-pressure anomaly accompanying a typhoon is caused by the centrifugal force drawing the air mass from the cyclone center. The right column of Fig. 5 shows the pressure anomalies during this period. It is found that the locations of dynamic dissipation and pressure anomaly are always very close along the path of the typhoon. However, more detailed structures are shown in the distribution of dynamic dissipation, while only a solid circle is found in the pressure anomaly field.

By comparing Figs. 2 and 5, it is found that at the incipient stage of a typhoon near Taiwan, dynamic

![Fig. 5. Daily evolution maps of (left) dynamic dissipation (0.01 W m⁻² K⁻¹), (middle) vertical shear of zonal wind (850 minus 200 hPa; m s⁻¹), and (right) pressure anomaly deviated from area mean (hPa) during 3–7 Aug 2005.](image-url)
dissipation is very weak but thermal dissipation is strong. This is because the rotational dynamic structure of a typhoon is not fully established in the early stages. As the typhoon moves northward, dynamic dissipation gradually becomes stronger but thermal dissipation becomes weaker. Near the eastern coast of China, dynamic dissipation becomes dominant. The typhoon involves complicated physical processes. Both dynamic and thermal dissipations play important roles in the evolution of a typhoon, but at different stages, one kind of dissipation might be dominant over the other.

It is seen from Fig. 5 that neither the wind shear nor the pressure anomaly can describe the dynamic structures of both the monsoon and typhoon at the same time, but the dynamic dissipation can. Therefore, non-equilibrium thermodynamics provides a more comprehensive description of the monsoon system. The strong seasonal change of wind is the second distinguishing feature for the South Asian monsoon. Based on Figs. 4 and 5, the monsoon feature of wind is properly addressed by dynamic dissipation. Therefore, the dissipation theory provides a reliable measurement of the South Asian monsoon, with two main features being described in a consistent way:

$$\sigma_L = J_w \frac{\partial}{\partial x_l} T + J_v \frac{\partial}{\partial x_l} T$$

represents the internal entropy production due to the large-scale eddy. Li and Chylek (2012) showed that the contribution in entropy production by the large-scale eddy appears mostly in midlatitude regions. The large-scale eddy plays a limited role in the South Asian monsoon since the large-scale eddy is very weak in the lower-latitude regions owing to less temporal variation in physical variables. In the left column of Fig. 6, it is shown that the contribution of a large-scale eddy is mostly in the higher-latitude regions, like the Tibetan Plateau and Turkish–Iranian Plateau. There are contours of dissipation by the large-scale eddy shown over both the Arabian Sea and the Bay of Bengal, especially in May and June. However, the magnitude of internal entropy production for large-scale eddy in the two monsoon branches is about one order smaller than that of the thermal dissipation shown in Fig. 1 (note that the plot scales are different in Figs. 1 and 6).

The rest of the terms in (3) show the physical processes of internal entropy production due to moisture flow diffusion and phase change:

$$\sigma_m = -R_v \frac{J_w}{T} \frac{\partial \ln T}{\partial x_l} - R_v \frac{J_v}{T} \frac{\partial \ln T}{\partial z} + \rho \frac{\partial (C - E)}{z} \ln T.$$

The first two terms are the typical dissipation process, as a large-scale-eddy moistening flow \((J_w/\lambda)\), and a small-scale moisture flux \([J_v/\lambda(T)] \) moving along the downgradient of \(\ln T\). This is very similar to the thermal flow moving along the downgradient of temperature. This is also an irreversible process and results in a positive internal entropy production. Therefore, the positive value of internal entropy production due to moisture flow diffusion represents the downgradient movement of eddy moisture flow and small-scale moisture flux in \(\tilde{\Pi}\). On the other hand, the negative value of internal entropy production indicates that the moisture flow is forced to move from a lower-\(\tilde{\Pi}\) region to a higher-\(\tilde{\Pi}\) region.

The last term in \(\sigma_m\) represents the irreversible process of condensation and evaporation. It is shown that evaporation \((-E \ln T\)) always corresponds to zero internal entropy production under saturated conditions \((\Pi = 1)\) and corresponds to positive internal entropy production under unsaturated conditions \((\Pi < 1)\) since the evaporation under nonsaturated condition is irreversible. Condensation \(C\) generally occurs when \(\Pi > 1\); hence, it corresponds to positive internal entropy production. Condensation generally occurs at \(\tilde{\Pi}\) of slightly over 1, and thus the value of \(C \ln \tilde{\Pi}\) is positive but extremely small. Therefore the contribution due to phase change is much smaller than that due to moisture-flow diffusion. Pauluis and Held (2002) was the first work that analyzed the role of moisture in atmospheric entropy.

It is found that the evolution of \(\sigma_m\) is also associated with the development of the monsoon. However, the value of \(\sigma_m\) is relatively small compared to the thermal and dynamic dissipation. We do not present the results of \(\sigma_m\) here.

4. Boundary entropy supply

From (2), the net boundary supply of entropy flux to the atmosphere is:

$$\left( \frac{\overline{F}_h \overline{F}_l + \overline{F}_c - \overline{R}_h \overline{F}_l + \overline{F}_c}{\overline{T}} \ln \overline{T} - \rho \overline{u} \alpha \overline{T}^3 \right)_{z=0}.$$

The large-scale eddy flows and the convective flux are supposed to be zero at the surface \((z = 0)\).

The left column of Fig. 6 shows the distributions of net boundary supply of the 10-yr climatological-mean entropy flux in the Asian monsoon region for 6 months, as in Fig. 1. The surface supplied entropy flux is due to surface sensible, latent heat; surface evaporation under nonsaturation condition; and surface stress, with the last two terms being relatively very small. In comparison
with the left column of Fig. 1, the pattern of boundary entropy supply is very similar to that of the internal entropy production for each month in both the summer and winter seasons. In June, July, and August, contours are easily seen in the two main monsoon branch regions in the Arabian Sea and the Bay of Bengal. These are related to the onset and progress of the South Asian monsoon. In September, contours of surface entropy supply are withdrawn from India over the Arabian Sea and the Bay of Bengal. Additionally, the big contour over the equator in the Indian Ocean retreats to the Southern Hemisphere.

FIG. 6. (left) Distribution of vertically integrated internal entropy production by large-scale eddies for 6 months, \((0.01 \text{ W m}^{-2} \text{ K}^{-1})\). (middle) Distribution of surface entropy supply \((0.1 \text{ W m}^{-2} \text{ K}^{-1})\). (right) Distribution of OLR \((\text{W m}^{-2})\).
The positive surface entropy flux corresponds to the surface supply of upward diabatic heat flows. The up-rising heat flows along the temperature gradient produces the internal entropy as analyzed in the previous section. Therefore, the distributions of external entropy supply and internal entropy production are correlated to some extent.

Outgoing longwave radiation (OLR) was used to detect the onset of monsoon (e.g., Soman and Slingo 1997). In the right column of Fig. 6 the corresponding results of OLR are shown. It is found that the large value contours of OLR occur mostly in western Asia near the African continent. This is seen to have little correlation to the summer monsoon structure. However, over the Arabian Sea and the Bay of Bengal, OLR presents patterns similar to those of the thermal dissipation shown in Fig. 1. OLR is strongly correlated to high cloud since the high cloud can shade the longwave radiation from reaching outer space. In the dissipation process, upward moist convective flow brings moisture into the upper troposphere, thereby forming high clouds. Therefore, larger thermal dissipation and the lower OLR are physically consistent.

In the monsoon onset regions, both the internal and external entropy productions are usually positive. Therefore, there must be some other processes to balance the entropy production locally. These processes are mainly the divergence of the large-scale eddy, entropy advection, and radiative sink. If we sum all the contributions of internal/external entropy production, large-scale eddy divergence, and entropy advection, then the total result is close to zero on a global average [see details in Li and Chylek (2012)]. Locally, the result cannot be zero with the residue representing the tendency of entropy.

The right column of Fig. 3 shows the daily evolution of the surface entropy supply during 3–7 August 2005. It is seen that the distributions of the surface entropy supply and latent–sensible dissipation are very similar over the Arabian Sea and the Bay of Bengal. This is because the surface convective flux is very small and the latent–sensible flux dominates. Additionally, the track of the typhoon is clearly shown in the surface entropy supply. This is similar to the situation for an extratropical cyclone (Li et al. 2014). We do not show the daily results of OLR since the distribution is very messy and no signal of a typhoon is found.

5. Regional three-dimensional diagnoses

Internal entropy production can describe not only the onset and strength of the monsoon but also its three-dimensional structure. Figure 7 shows the 10-yr climatological-mean internal entropy production at different heights in July. We focus on the Indian subcontinent from 5° to 35°N and from 65° to 90°E.

The left column of Fig. 7 shows the results of internal entropy production. At $\eta = 0.925$, there are big contours near the western coast of India. Along the western coast of India, air rising by the topographic barrier of the mountainous Ghats causes large internal entropy production as a result of strong vertical dissipation. Inside India, the dissipation is larger in the southwestern part compared to that in the southeastern part. The internal entropy production over the Himalayan region is also large. As mentioned above, in the summer season the Tibetan Plateau is a source of heat for the atmosphere. The resulting motion of rising air dissipates the vertical temperature gradient and produces internal entropy production. In the northern part of India, the internal entropy production is large in the Himalayan region and becomes smaller in Rajasthan. This indicates weaker thermal dissipation there. Rajasthan region (green circle) is known as the Great Indian Desert, though this region is very close to ocean. The weak thermal dissipation, and the consequent lower upward thermal flow, can help us to understand the dry status in the Rajasthan region.

At $\eta = 0.7$ the internal entropy production becomes weaker over India and the Arabian Sea. The result, however, does not change much over the Himalayas. The contours near the western coast are shifted toward India in contrast with that of $\eta = 0.925$. At $\eta = 0.5$, the distribution of internal entropy production is similar to that of $\eta = 0.7$. In both $\eta = 0.7$ and $\eta = 0.5$, the region of Rajasthan is very different from its surroundings, with an obvious weaker internal entropy production.

To understand the results of internal entropy production in Fig. 7, we also present the corresponding distributions of large-scale eddy flow and small-scale convective flux. The vertical component of large-scale moistening flow $J_{\alpha3} = \lambda \rho \bar{q} \bar{v}$ is shown in the middle column of Fig. 7 for three different heights, where $\omega$ is the pressure vertical velocity. Also, the result of small-scale convective flux $F_{s}$ is shown in the right column. The role of $J_{\alpha3}$ and $F_{s}$ in dissipation is shown in (3).

It is found the large-scale eddy moistening flow is generally upward in northern India and downward in southern India. The ascending thermal eddy flow in the Himalayan region spreads out southward and gradually sinks in the lower-latitude areas. It is found that the eddy moistening flow is strongly downward in Rajasthan, especially at the lower level of 0.925. The downward eddy moistening flow suppresses the upward motion of moisture and limits the precipitation as the moisture cannot reach supersaturation status.

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FIG. 7. (left) Distribution of internal entropy production at different heights ($10^{-6}\text{Wm}^{-2}\text{K}^{-1}$). (middle) Distribution of vertical components of large-scale-eddy moistening flow at different heights and (right) distribution of convective flux at different heights ($\text{Wm}^{-2}$). The results are for 10-yr-mean July.
The small-scale convective flux is mostly upward in the considered regions. The convective flux generally increases with height, since the convective condensation usually occurs in the middle troposphere only when the convergence of the moisture flux becomes positive there. It is clearly shown that the upward convective flux is very weak in Rajasthan, which implies a stable condition and small internal entropy production. This is visible in the left column of Fig. 7.

6. Conclusions

The primary objective of this study is to investigate atmospheric entropy structure and evolution characteristics in association with the seasonal progression of the South Asian monsoon. The monsoon is a climate system that consists of well-organized large-scale thermal and dynamic structures. The dissipation to the orderliness of the system naturally occurs in the monsoon development stages. This study shows that the internal entropy production (dissipation) is usually associated with monsoon development. In the Arabian Sea and the Bay of Bengal, the onset and withdrawal of the dissipation contours during the summer monsoon are clearly shown. It is found that the thermal dissipation is mainly caused by the upward convective heat flux and, as a result, the thermal dissipation is highly correlated to precipitation. The structure and evolution characteristics of a monsoon can be, to a certain extent, addressed by dynamic dissipation patterns. In June, July, and August, the dynamic dissipation is very strong over the Arabian Sea and southwestern India. It is shown that dynamic dissipation can describe the monsoon dynamic structure more clearly than the wind shear pattern. This difference is particularly evident over the Indian subcontinent. The strong seasonal changes of precipitation and wind are distinguishing features for the South Asian monsoon. Both features can be properly described by internal entropy production in a consistent way.

In addition to climatological monthly evolution characteristics, the daily evolution features of thermal and dynamic dissipation are demonstrated during 3–7 August 2005. During this period, a typhoon in the western Pacific was developing. In the Arabian Sea and the Bay of Bengal, the thermal dissipation is found in strong correlation to precipitation. In addition, the track of the typhoon in the early stage is clearly shown in the thermal dissipation. It is found that the dynamic dissipation can provide a unified picture that describes both the large-scale monsoon pattern and smaller-scale typhoon structure at the same time. This cannot be achieved with the use of a traditional method that analyzes conventional variables such as wind shear, sea level pressure, and precipitation.

It is shown that the correlation between the surface external entropy supply and the internal entropy production is strong over the ocean. Therefore, the monsoon structures can also be explored by the external entropy supply.

This study is based on reanalysis data. In climate model simulations, the principle of maximum internal entropy prediction (Li 2009) probably can be used to restrain the monsoon development.

The methodology of the atmospheric entropy analysis demonstrated here provides a new way to describe the developing characteristics of a monsoon. The thermal dissipation, dynamic dissipation, and external entropy supply clearly reveal distinctive physical processes occurring at different development stages of the monsoon. The physical insight derived from such an analysis is a useful extension to the traditional methods.

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