The Dissipation Structure of Extratropical Cyclones

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ABSTRACT

The physical characteristics of extratropical cyclones are investigated based on nonequilibrium thermodynamics. Nonequilibrium thermodynamics, using entropy as its main tool, has been widely used in many scientific fields. The entropy balance equation contains two parts: the internal entropy production corresponds to dissipation and the external entropy production corresponds to boundary entropy supply. It is shown that dissipation is always present in a cyclone and the dissipation center is not always coincident with the low-pressure center, especially for incipient cyclones. The different components of internal entropy production correspond to different dissipation processes. Usually the thermal dissipation due to turbulent vertical diffusion and convection lags geographically the dynamic dissipation due to wind stress. At the incipient stage, the dissipation is mainly thermal in nature. A concept of temperature shear is introduced as the result of thermal dissipation. The temperature shear provides a useful diagnostic for extratropical cyclone identification. The boundary entropy supply and the entropy advection are also strongly associated with cyclones. The entropy advection is generally positive (negative) in the leading (trailing) part of a cyclone. A regional study in the western Pacific clearly demonstrates that the surface entropy flux and temperature shear are the most reliable early signals of cyclones in the cyclogenesis stage.

1. Introduction

The genesis and evolution of cyclones has received considerable attention for almost a century. In the early twentieth century, Bjerknes (1919) proposed the frontal-cyclone model. Since that time, cyclones have been the subject of ongoing research aimed at furthering our understanding and prediction of cyclones, especially for extratropical marine cyclones.

In the early stage of research the simplified quasi-geostrophic equations (Charney 1947) were used for studying the evolution of cyclone waves from genesis to decay. Usually the idealized midlatitude flow is represented by the basic states, which are zonally uniform but possess vertical and horizontal shears. Studies of cyclone-wave-scale disturbances to the basic states were in the linear regions of growth (Pedlosky 1971; Mudrick 1974) and in extension to the nonlinear regions (Simons 1972; Simmons and Hoskins 1978).

Moist dynamics related to cyclones can be explored from an energetics point of view. The release of latent heat plays an important role in energetics. Compared to the dry case, moist cyclones possess very similar, but enhanced, growth and decay rates during their life cycles. The transport of heat and momentum fluxes is strengthened as well. The eddy kinetic energy is also enhanced owing to condensation (Emanuel et al. 1987; Orlanski and Katzfey 1991; Neiman and Shapiro 1993; Neiman et al. 1993; Balasubramanian and Yau 1996; and others).
A generally accepted theoretical explanation of cyclogenesis is based on baroclinic instability. A small initial disturbance grows exponentially as a result of the unstable basic state (Charney 1947; Eady 1949; Hoskins and West 1979; Plumb 1986; and others). The generalization of baroclinic instability to nonmodal growth has also been developed.

In the analysis of the Lorentz energy cycle, the extratropical cyclones acts as a mechanism for converting potential energy that is created by pole-to-equator temperature gradients to eddy kinetic energy (Holton 2004). There also have been studies based on potential vorticity and wave activity budgets (Davis and Emanuel 1991; Davis 1992; Black and Dole 1993; Sinclair 1993; Orlanski 2003; and others).

In the present work, an effort is made to understand the physics of extratropical cyclones using a new approach based on nonequilibrium thermodynamics. Nonequilibrium thermodynamics (e.g., Glansdorff and Prigogin 1971; deGroot and Mazur 1984; Hu 2002), using entropy as its main tool, has been widely used in many fields in physics, chemistry, and biology. In the last several decades, the entropy related to the second law of thermodynamics has also been applied to general circulation of the atmosphere (Peixoto and Oort 1992; Li et al. 1994; Johnson 1997; Goody 2000; Takamitsu and Kleidon 2005; Fraedrich and Lunkeit 2008; Pascale and Gregory 2009; Lucarini et al. 2011), moist convection (Paulus and Held 2002), and synoptic-scale severe atmospheric systems (Liu and Liu 2004). Additionally, Nicolis and Nicolis (1980) and Pujol and Llebot (1999) have applied the second differential of entropy to instability in low-dimensional climate models. A significant effort has also been to study the maximum entropy production principle (MEPP) using simple climate models (Paltridge 1975; O’Brien and Stephens 1995; Lorenz et al. 2001; Ou 2001; Okawa et al. 2003; Shimokawa and Ozawa 2007; Wang et al. 2008; Li 2009; Lucarini 2009; Lucarini et al. 2010; and many others). Li (2009), Lucarini (2009), and Lucarini et al. (2009) have critically discussed MEPP. However, the application of atmospheric entropy to detailed weather/climate phenomena has seldom been investigated. In any system, irreversible processes are always accompanied by positive internal entropy production, which dissipates the orderliness of the system. The concept of dissipation was first proposed by Prigogine (1969). The analysis of the internal entropy production (dissipation) and the external entropy production (boundary entropy supply) is always essential for exploring the change of orderly structure in any complicated system.

As mentioned above, the baroclinic process is generally the forcing for extratropical cyclones because they form along zones of temperature and moisture gradient known as frontal zones. Cyclones generally contain strong thermal, dynamic, and moisture structures and are accompanied by heavy precipitation. For a developing cyclone, the dissipation to orderliness of the system is generally different from that of its surroundings. Thus, the dissipation structure can be indicative of cyclones. Therefore, the atmospheric entropy provides a new way to analyze the physical structure of cyclones. The purpose of this study is to understand the evolution of cyclones with the aid of nonequilibrium thermodynamics based on the analysis of atmospheric entropy production. In section 2, the basic theory of atmospheric entropy is introduced. In section 3, the internal entropy production related to cyclone is discussed. Section 4 shows the relation between the surface entropy supply and cyclones. Section 5 focuses on a regional case study for cyclogenesis. In section 6, we conclude with a brief summary.

2. Theoretical background for atmospheric entropy

Let \( s \) be the entropy per unit mass. The balance equation for \( s \) is (Emanuel 1994)

\[
    s = c_{pd} \ln \theta + \frac{\lambda q_v}{T} - R_v q_v \ln \mathcal{H},
\]

where \( c_{pd} \) is the specific heat at constant pressure for air, \( T \) is the temperature, \( \theta \) is the potential temperature, \( q_v \) is the specific humidity, \( \lambda \) is the latent heat of evaporation of water, \( R_v \) is the gas constant for water vapor, and \( \mathcal{H} \) is the relative humidity. From (1), we obtain

\[
    \frac{ds}{dt} = \frac{c_{pd} d\theta}{\theta} + \frac{\lambda dq_v}{T} - R_v \ln \mathcal{H} \frac{dq_v}{dt},
\]

where \( d/dt \) is the Lagrange derivative. Terms of approximately \( 1/T^2 \) have been neglected in (2), since they are two orders smaller. The term related to \( d \ln \mathcal{H}/dt \) is found to be much smaller than other terms. By the first law of thermodynamics,

\[
    T \frac{d\theta}{dt} = Q_r + Q_h + Q_f + Q_1 + \frac{\lambda}{c_{pd}} (C - E),
\]

where \( Q_r, Q_h, \) and \( Q_f \) are diabatic heating rates corresponding to radiation, sensible heating, and frictional heating, respectively; \( Q_1 \) is the heating rate corresponding to convection (deep and shallow); and \( C \) and \( E \) are the condensation and evaporation rates, respectively. In National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) data (Kistler et al. 2001), the sensible heating rate corresponds to the turbulent vertical diffusion process, and \( (\lambda/c_{pd})(C - E) \) is the resolved large-scale heating rate corresponding to the net condensation.
The reanalysis data provide only the diabatic heating rates. The procedure for obtaining the vertical diabatic heat flux from the diabatic heating rate is shown (Li and Chylek 2012). Written in flux form, (3) becomes

\[
\frac{T}{\theta} dt = -\frac{1}{c_{pd} \theta} \left( \frac{\partial F_h}{\partial z} - \rho \tau_{3, \alpha} \frac{\partial u_{\alpha}}{\partial z} \right) + Q_1 + \lambda \cdot (C - E) + Q_r, \tag{4}
\]

where \( F_h \) is the sensible heat flux (vertical diffusion heat flux) and \( \tau_{ij} \) is the diffusion tensor. Subscript \( i \) (or \( j \)) takes values 1, 2, and 3, representing two horizontal and one vertical directions, as \( x_1, x_2, x_3 = x, y, z \) and \( u_1, u_2, u_3 = u, v, w \) are velocities in the \( x, y, z \) directions. We sum the repeated indices \( \alpha \) from 1 to 2 as \( \tau_{3, \alpha} \frac{\partial u_{\alpha}}{\partial z} = \tau_{3, 1} \frac{\partial u_1}{\partial z} + \tau_{3, 2} \frac{\partial u_2}{\partial z}. \)

Similarly, as shown in Li and Chylek (2012), the balance equation for \( q_v \) is

\[
\frac{dq_v}{dt} = \frac{c_{pd} Q_1}{\lambda} - \frac{c_{pd} Q_2}{\lambda} + E - C = -\frac{1}{\lambda \rho} \frac{\partial F_l}{\partial z} - \frac{c_{pd} Q_2}{\lambda} + E - C, \tag{5}
\]

where \( Q_1 \) is latent heat due to vertical turbulent diffusion, \( F_l \) is the corresponding latent heat flux, \( Q_2 \) is the moistening heating in deep/shallow convection, and \( Q_2 / \lambda \) is the corresponding moistening rate.

Generally, the convective heating of \( Q_1 \), as the divergence of convective heat flux, and the convective moistening heating of \( -Q_2 \), as the divergence of moistening flux, do not occur in the same location. The convective condensation usually occurs in the middle troposphere only when the convergence of convective heat flux becomes positive. In this process, part of condensed liquid/ice water will fall down as precipitation and drizzle, while the convective moistening process usually produces cooling and occurs in the lower atmosphere (Yanai et al. 1973). The difference \( Q_1 - Q_2 = \frac{-1}{\lambda} \frac{\partial}{\partial z} \left\{ \frac{\rho}{\lambda} \left[ (w^* h_v^*) \right] \right\} \), where the asterisk represents the deviation from the grid-scale mean of the variables in the square brackets and \( h_v \) is moisture static energy (Emanuel 1994). We denote \( F_c = \rho [w^* h_v^*] \) as the convective heat flux.

From (2), (4), and (5), and using the continuity condition, the entropy balance equation is derived:

\[
\frac{d\sigma}{dt} = \frac{\partial \sigma}{\partial t} + \frac{\partial \rho u_s \sigma}{\partial x} = \frac{\partial}{\partial z} \left( \frac{F_h + F_l + F_c}{T} \right) - \frac{\rho u_{\alpha} \tau_{3, \alpha}}{T} - R_v \frac{F_l - F_{cm}}{\lambda \ln \mathcal{H}} + \sigma + \frac{c_{pd} \rho Q_r}{T}, \tag{6}
\]

where \( \sigma \) is the internal entropy production, and

\[
\sigma = \left( F_h + F_l + F_c \right) \frac{1}{T} - \rho u_{\alpha} \frac{\partial}{\partial z} \left( \frac{\tau_{3, \alpha}}{T} \right) - R_v \frac{F_l - F_{cm}}{\lambda \ln \mathcal{H}} + R_v \rho (C - E) \ln \mathcal{H}, \tag{7}
\]

where \( F_{cm}/\lambda \) is the convective moistening flux corresponding to \( Q_2 \) [i.e., \( Q_2 = \frac{1}{\lambda} (c_{pd} \rho) (\partial F_{cm}/\partial z) \)]. On the right-hand side of (7), the first term is the internal entropy production due to the diabatic heat fluxes, the second term is due to wind stress, and the third and fourth terms are due to water vapor diffusion and phase change, respectively.

The balance equation of (6) is as in (3) in Li and Chylek (2012), but derived without considering the large-scale eddy. The large-scale eddy, coming from the temporal variation, only appears in climate studies with considerably longer time scale but does not apply to the events with short time scale such as cyclones. In the following we will focus on each physical process involved in (6) and (7). In NCEP–NCAR data, the diabatic heat fluxes within the atmosphere are not available but can be obtained from the diabatic heating rates as discussed in Li and Chylek (2012).

Equation (6) shows that the time evolution of entropy is caused by the divergence of external entropy flux, internal entropy production, and an entropy sink due to radiation, since generally \( c_{pd} \rho Q_2 / T < 0 \) 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. The atmospheric temperature gradient is mostly created by the non-homogeneous distribution of radiative heating. Regionally, the air–sea interactions at the western boundary currents also help maintain the atmospheric temperature structure owing to strong ocean fronts and large instability. However, for a detailed weather event with a very short time scale such as a cyclone, radiation plays a very minor role in comparison to the other thermal and dynamic factors. This is because the radiative heat, absorbed by the cyclone, is about two orders smaller than the sensible–latent–convective heat supplied to the cyclone. It is found the including of radiation has negligible impact on the structures of cyclone.

### 3. Internal entropy production related to dissipation of cyclones

The NCAR–NCEP reanalysis data are often used to study Northern Hemisphere extratropical cyclones. Froude et al. (2007) showed that the NCAR–NCEP ensemble prediction systems (EPS) are comparable to the European...
Centre for Medium-Range Weather Forecasts (ECMWF) EPS in predictive skill. Lucarini et al. (2007) compared 19 global climate models with the NCEP–NCAR and ECMWF reanalyses for eastward-propagating baroclinic waves. It is found that only very-high-resolution global climate models can have a good agreement with the re-analysis result for traveling baroclinic waves.

Generally, in January, there are many extratropical cyclones in the North Pacific. Here we choose January 1995 (result is similar for January in other years). The top-left panel of Fig. 1 displays the distribution of the surface pressure anomaly at 0000 UTC 7 January. We choose 7 January because some cyclones that appeared in the Pacific on this day had undergone cyclogenesis in previous days of the same January. Such cyclogenesis process will be studied in section 5. The surface pressure anomalies of cyclones have widely been used to detect cyclone presence. Surface pressure anomaly is defined as the surface pressure difference between a certain time in January and the 50-yr January mean from 1951 to 2000. At 0000 UTC, there are three low-pressure centers in the North Pacific and another low-pressure center developing off the coast of Asia near Japan. To track the evolution of the cyclones, a number is assigned to each cyclone in the North Pacific.

At 1200 UTC 9 January, 2.5 days later, the number-1 and -2 low-pressure centers disappeared. The number-3 low-pressure center, which was close to Japan on 7 January, has moved to close to the western coast of North America. An evolution of a cyclone is therefore clearly displayed. Figure 1 shows that the pressure anomaly can describe to some extent the development and movement of cyclones. Using pressure anomaly or geopotential height to study cyclones is a very common approach (e.g., Lambert 1995). The low-pressure anomaly accompanying a cyclone is caused by geostrophic balance, as the pressure gradient force (from the center of a cyclone to its outside) and the Coriolis force must be in balance, as the pressure gradient force (from the center of a cyclone to its outside) and the Coriolis force must be in balance. The low-pressure anomaly method is not effective in the case with weak cyclic rotation. For example, there is no low-pressure anomaly appearing over the Atlantic basin because of the weak cyclic structures for the existing cyclones (or called storms).

Ertel’s potential vorticity (PV) has been widely applied to analyze cyclogenesis and cyclone development (e.g., Davis and Emanuel 1991). In hydrostatic pressure coordinates,

\[
PV = -g \left[ \left( \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) \frac{\partial \theta}{\partial p} - \frac{\partial u}{\partial p} \frac{\partial \theta}{\partial y} + \frac{\partial v}{\partial p} \frac{\partial \theta}{\partial x} \right],
\]

where \(g\) is the gravitation constant, \(p\) is the pressure, and \(f\) is the Coriolis parameter.

The middle row of Fig. 1 shows the corresponding distributions of PV at 372 hPa. (It will be shown in Fig. 3 that PV can provide clear information of a cyclone in a higher-altitude region.) The result at 0000 UTC 7 January clearly shows the tracks of cyclones as measured by PV. For a same cyclone, it is found that the center of the low-pressure anomaly and the center of PV are generally close but not exactly collocated. During the incipient stage of a cyclone, the structure must tilt westward with height for it to grow and, hence, PV in the upper troposphere is not well collocated with the cyclone in the lower atmosphere. Later in its life cycle, when the cyclone is mature, it becomes more vertically stacked and, thus, the center of PV becomes more vertically aligned with the cyclone center. At 1200 UTC 9 January, the number-3 center of PV has moved to close to the western coast of North America. Overall, it is found that the centers of PV and low-pressure anomaly are close, but the location of PV usually lags behind the location of pressure anomaly.

Based on energetics, the release of latent heat plays an important role for evolution of a cyclone. Let us consider the diabatic heating rate, which includes the sensible and latent heat due to vertical turbulent diffusion, moistening convective heat, and the fractional heat due to wind stress. In the bottom panels of Fig. 1, the distributions of the vertically integrated diabatic heating rate are shown. At 0000 UTC 7 January, it is found in the North Pacific that there are three centers of distinguished diabatic heating rate with locations close to centers of low surface pressure. Over the Atlantic basin, there is a region with large diabatic heating rate, which indicates the existence of a storm over the Atlantic basin. This will be supported by following other analysis methods.

The existence of a cyclone can be also evident from the precipitation point of view (not shown). In the Pacific, the center of precipitation is usually located in the leading edge of a cyclone. The precipitated water at the leading edge of the cyclone is replenished through the convergence of water vapor (Neiman and Shapiro 1993). A cyclone usually has complicated structure. There are fronts associated with cyclones—with warm and cold sectors and warm and cold conveyor belts. Precipitation generally occurs along the fronts. In the Atlantic, there is a heavy precipitation zone at 0000 UTC 7 January, which is roughly consistent with the location of the higher value of diabatic heating rate. This displays the presence of a storm there.

Figure 2 shows the distributions of vertically integrated internal entropy production based on (7). The internal entropy production reveals an irreversible dissipation process. A typical form of internal entropy production in (7) is
FIG. 1. (top) The distributions of surface pressure anomaly at 0000 UTC 7 Jan and 1200 UTC 9 Jan, with contour interval of 5 hPa. (middle) The distributions of PV at 372 hPa with contour interval of $2 \times 10^{-8}$ kg m$^{-1}$ s$^{-1}$ K. (bottom) The distributions of vertically integrated diabatic heating rate with contour interval of 0.015 W m$^{-2}$. Dashed lines are negative values.
Fig. 2. The distributions of vertically integrated internal entropy production from 0000 UTC 7 Jan to 1200 UTC 9 Jan, with contour interval of 0.025 W m$^{-2}$ K$^{-1}$.
\[ \sigma = F \frac{\partial}{\partial z} \frac{1}{T} \]  

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

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

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

\[ \sigma = k \frac{1}{T^2} \left( \frac{\partial T}{\partial z} \right)^2 > 0. \]  

Equation (9) shows that the thermal flow moves along the gradient of temperature from the higher-temperature region to the 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, which results in a reduction of the temperature gradient. Therefore, the positive internal entropy production is referred to as dissipation (Prigogine 1969). In Lucarini (2009), the degree of irreversibility is introduced to quantify the relative relevance of the irreversible process versus the dissipation of kinetic energy.

We would like to emphasize that Fourier’s law in (9) is generally not followed by the latent heat flux and convective heat flux. However, if the sensible–latent–convective heat flux, \( F_s + F_l + F_c \), is upward, thereby downgradient to the vertical temperature, the dissipation process discussed above is held true and the positive internal entropy production is related by

\[ \sigma = - \frac{F_s + F_l + F_c}{T^2} \frac{\partial T}{\partial z} > 0, \]  

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

In the cyclone case, an important ingredient for cyclone development is the baroclinicity due to north–south temperature difference. Even by a simple Eady model, baroclinicity can generate eastward baroclinic waves near the lower boundary (Eady 1949). The cyclonic flow brings a warm and moist front poleward over the leading part of a cyclone and a cold and dry front equatorward over the trailing part. Relative to the equatorward cold front, the poleward warm front progresses slowly as cooler air ahead of the system is denser and therefore more difficult to dislodge. As occlusion occurs, the warm air mass is pushed upward over the cold air mass (Wallace and Hobbs 2006). The detailed explanation for the trough of warm air aloft (“trowal” for short) can be found in Martin (1998, 1999). Thus, the center of gravity is lowered when a warmer (lighter) parcel is moved higher. Following the Lorenz energy circle, potential energy is extracted from the system and converted to kinetic energy. This strengthens the dynamic (wind) structure of a cyclone. At the same time, the upward movement of a warmer air parcel is downgradient to the vertical temperature, which corresponds to the dissipation as discussed above. Detailed calculations show that the thermal flow of \( F_h + F_l + F_c \) is generally upward inside a cyclone. Therefore, the driving of cyclonic flow by baroclinicity, the Lorenz energy circle, and the dissipation are physically consistent.

Therefore, dissipation distinguishes the cyclone from its surroundings, and the dissipation structure can be indicative of cyclones. In comparison with the distributions of the surface pressure, PV, and diabatic heating, Fig. 2 shows that the distributions of internal entropy production can reveal the details of the cyclones more clearly.

Besides 0000 UTC 7 January and 1200 UTC 9 January, all the detailed results within the subsequent 2.5 days are shown, in that the evolution of cyclone is clearly displayed. For example, the number-3 cyclone was in the western Pacific at 0000 UTC 7 January, then it gradually moved eastward and reached the western coast of North America at 1200 UTC 9 January. Over the Atlantic basin, the signals of storm are clearly shown based on the dissipation structure, which is consistent with the results of diabatic heating.

The internal entropy production can describe the strength of a cyclone, show its three-dimensional structure and provide information on its track. In the top row of Fig. 3, the two-dimensional dissipation structures between 120°E and 120°W are displayed for two moments of 0000 UTC 7 January and 1200 UTC 9 January. The displayed dissipations show the meridionally averaged results from 31° to 43°N, a region, which is frequented by cyclones. By comparing the results at the two moments, the development and tracks of the cyclones toward the east are clearly evident. In the middle row of Fig. 3, the vertical distributions of the meridionally averaged PV in the same zone are shown. It is found that PV has a strong structure above 500 hPa. It is known that the baroclinic forcing in the lower atmosphere generally has a profound effect on disturbances in the upper layers (e.g., Orlanski 2003). In comparison with the top row, it can be seen that the longitudinal locations of PV and internal entropy production (dissipation) are close to some extent.

In the bottom row of Fig. 3, the vertical distributions of the meridionally averaged diabatic heating rate are shown. It is found that the diabatic heating has a strong
structure in the lower atmosphere, similar to dissipation. Generally, a cyclone shows negative heating in its leading edge. This is especially apparent for mature cyclones in the eastern Pacific. The negative heating rate in the leading edge is mostly due to convection. The convective moistening process usually produces cooling and occurs in the lower atmosphere (Yanai et al. 1973).

In Fig. 3, although the energetic method can be used to describe the structure of cyclone as well, results are usually less clear compared to that obtained by the internal entropy production.

To understand the detailed physics of dissipation, in the following we discuss the contributions from different physical processes. In Fig. 4, the top row displays the
FIG. 4. (top) The distributions of the vertically integrated internal entropy production corresponding to thermal dissipation with contour interval of 0.025 W m\(^{-2}\) K\(^{-1}\), (middle) the meridionally averaged (31°–43°N) cross section of thermal dissipation with contour interval of 10\(^{-2}\) W m\(^{-3}\) K\(^{-1}\), and (bottom) the vertical temperature distributions in the same zone with contour interval of 2 K.
contribution of the vertically integrated internal entropy production due to the turbulent vertical diffusion and convection as

$$\sigma_{i} = (F_{h} + F_{l} + F_{c}) \frac{1}{\bar{z}} T. \quad (10)$$

We term the internal entropy production by turbulent vertical diffusion and convection as thermal dissipation. In the incipient stage of a cyclone, moving off the eastern coast of Asia near Japan, thermal dissipation is relatively strong. As the cyclone moves eastward, the track of thermal dissipation is close to that of the total dissipation shown in Fig. 2. It is interesting that there are large contours of thermal dissipation in the Atlantic basin, which is consistent with the distributions of diabatic heating. This means that storms in the Atlantic basin are dominated by a thermal structure.

The meridionally averaged vertical distributions of thermal dissipation are displayed in the middle row of Fig. 4 for the same zone as in Fig. 3. It is shown that the vertical structures of cyclones are well represented through thermal dissipation. The movement of cyclones and their change of strength are shown clearly. Thermal dissipation has extended into the middle troposphere owing to the contribution of convection. In the atmosphere, much of the latent heat release occurs in the middle troposphere instead of in the boundary layer (Emanuel 1994).

Based on (10), thermal dissipation displays a physical process that the sensible, latent, and convective flows move downgradient to temperature, which causes a decrease of temperature gradient. In the bottom row of Fig. 4, the meridionally averaged temperature distributions are displayed for the same zone as in the middle row. It is found that the contour lines of temperature are always lower in a cyclone-located region. Thus, the temperature gradient in such cyclone is larger than its surroundings. This means a higher orderly structure in temperature for the cyclone. During the dissipation process, cyclones transport heat and reduce the temperature gradient. Thus, the temperature structure has been dissipated, as it is shown that contour lines of temperature move up after a cyclone has passed. Consequently, the temperature structure is a clear indication for the existence of thermal dissipation.

A simple way to demonstrate the strong vertical temperature gradient in a cyclone-located region is through the temperature difference between an upper level and the surface (called temperature shear). In Fig. 5, the distributions of temperature shear (between 500 hPa and surface) are shown for the same cases as in Fig. 2. A cyclone is always located close to a larger temperature shear region. Especially, for an incipient cyclone in the eastern Pacific near Japan, the signal of temperature shear is much clearer than the signal of pressure anomaly shown in Fig. 1.

It is shown in the bottom row of Fig. 4 that the contours of temperature can be distorted in a very high region. This is probably due to the influence of the large-scale planetary wave. It is found that the results in Fig. 5 are similar by choosing the upper-level temperature shear from 700 to 400 hPa. However, if the upper level is set too high, like 300 hPa and above, the contours of temperature shear can be less clear in collocation with the center of dissipation.

Since the temperature shear is clear evidence for dissipation, in Fig. 6 more detailed results of temperature shear are presented for some of the other days of January 1995. We only present the result in the North Pacific. In addition, the corresponding results of pressure anomaly and PV are included. The results PV are vertically integrated. The integration is from the surface to 200 hPa. From Fig. 3, the distributions of the vertically integrated PV are mostly dependent on the results of PV at upper troposphere. The distributions are similar for the vertically integrated PV and the PV at one level in the upper troposphere. At 0000 UTC 12 January, there is a low-pressure center in the eastern Pacific. The corresponding centers of temperature shear and PV are also shown there. In the western Pacific near Japan, the signal of temperature shear is shown by both the PV and temperature shear, but not shown by pressure anomaly. In the cyclogenesis stage, the pressure anomaly is not sufficient to be used to detect a cyclone (see section 5). Generally, in the eastern Pacific, where the cyclones become more mature, the centers of temperature shear are closer to the centers of pressure anomaly, in comparison with the results of PV.

In the western Pacific, the centers of temperature shear and PV are more collocated. In a few cases, the center of temperature shear is not close to the center of pressure anomaly. For instance, at 0000 UTC 28 January, the center of temperature shear is far behind the center of pressure anomaly. The different physical variables describe different physical processes. The centers of the low surface pressure, vorticity, and thermal dissipation are not expected to be always collocated.

The dissipation occurs not only to the temperature structure (i.e., temperature gradient) but also to the dynamic and moisture structures (details shown in the following). In Fig. 7, the top row displays the contribution of the vertically integrated internal entropy production due to the wind stress as

$$\sigma_{\alpha} = -\rho u_{\alpha} \frac{\partial}{\partial z} \left( \frac{T \alpha}{h} \right). \quad (11)$$
Fig. 5. The distributions of temperature shear between 500 hPa and surface from 0000 UTC 7 Jan to 1200 UTC 9 Jan, with contour interval of 1.5 K.
We term the internal entropy production by wind stress as “dynamic dissipation.” Dynamic dissipation shows the dissipation to dynamic structure of $\tau_{a,3}$ by horizontal wind $u_a$.

By comparing Figs. 4 and 7, it is found that at the incipient stage of a cyclone, moving off the eastern coast of Asia near Japan, dynamic dissipation is very weak, but thermal dissipation is strong. As the cyclone moves eastward, dynamic dissipation gradually becomes stronger. Near the eastern coast of North America, dynamic dissipation becomes dominant. The cyclone is a complicated physical process. Both dynamic and thermal dissipation play important roles in the evolution of a cyclone, but at different stages, one kind of dissipation might be dominant over the other.

In the Atlantic basin, where the thermal dissipation of storms is generally strong, the corresponding dynamic dissipation is too weak to be noticed. The weak dynamic dissipation is consistent with the result of the unnoticed low-pressure anomaly in the Atlantic basin, since the dynamic cyclic wind structure is very weak there.

The meridionally averaged vertical distributions of dynamic dissipation are displayed in the middle row for the same zone as in Fig. 3. It is shown that the vertical structures of cyclones are well represented through the dynamic dissipation. It is shown that dynamic dissipation generally occurs in the mature stage of a cyclone. In the mature stage a strong cyclic structure can support a large horizontal wind to dissipate the existing dynamic (wind stress) and temperature structures.

In a dynamic dissipation process, temperature plays a weak role, since in the lower atmosphere the vertical variation of $\tau_{a,3}$ is much larger than that of temperature. Therefore,

\[ \sigma_d = -\frac{\rho}{T} u_a \frac{\partial \tau_{a,3}}{\partial z} = -\frac{\rho}{T} u \frac{\partial \tau_{1,3}}{\partial z} - \frac{\rho}{T} v \frac{\partial \tau_{2,3}}{\partial z}. \]

Thus, the dynamic dissipation shows dissipation of the dynamic structure of wind stress by $u_a$. Since $\tau_{a,3}$ represents the horizontal momentum in the vertical direction, $-u_a \tau_{a,3}/\partial z$ shows the downgradient of horizontal momentum induced by the horizontal wind. This is similar to the physical process shown in (8). In the bottom row of Fig. 7, the meridionally averaged vertical distributions of wind stress $\tau_{2,3}$ are shown. This is the meridional
FIG. 7. (top) The distributions of the vertically integrated internal entropy production corresponding to dynamic dissipation with contour interval of 0.01 W m$^{-2}$ K$^{-1}$, (middle) the meridionally averaged (31$^\circ$–43$^\circ$N) cross section of the dynamic dissipation with contour interval of $10^5$ W m$^{-3}$ K$^{-1}$, and (bottom) the vertical distributions of $\tau_{2,3}/T$ in the same zone with contour interval 0.25 m$^2$ s$^{-2}$ K$^{-1}$. Negative values are shaded.
component of wind stress. It is interesting to find that \( \tau_{2,3} \) is negative in the leading part of a cyclone and becomes positive in its trailing part. Therefore, \(-\partial \tau_{2,3}/\partial z\) is positive in the leading part and negative in the trailing part, since wind stress generally decreases with height.

The airflow around a cyclone brings the northern wind (\( v > 0 \)) to the leading part of a cyclone and southern wind (\( v < 0 \)) to the trailing part. Both parts generate the positive internal entropy production by dissipation of the dynamical structure. The similar result is found for the zonal component of wind stress.

Figure 8 displays the contribution of internal entropy production due to moisture field as

\[
\sigma_m = -R_v \frac{F_i + F_{cm}/\lambda}{\lambda} \frac{\partial \ln \mathcal{H}}{\partial z} = -R_v \frac{1}{\lambda} \frac{F_i + F_{cm}}{\lambda} \frac{\partial \mathcal{H}}{\partial z},
\]  

(12)

Since \( F_i/\lambda \) and \( F_{cm}/\lambda \) are the moisture flow for moist turbulent diffusion and convection, respectively, we term the internal entropy production by moisture flows as "moisture dissipation." It is found that generally the tracks of eastward-moving cyclones in the North Pacific are not clearly demonstrated by moisture dissipation. The signal of \( \sigma_m \) is much weaker than those of \( \sigma_t \) and \( \sigma_d \). We therefore only demonstrate the meridionally averaged vertical distributions of moisture dissipation for the same zone as in Fig. 3. The dissipation structures are apparent for some cyclones, like the number-3 cyclone. The physics is similar for moisture dissipation and thermal dissipation, since (12) is analog to (10) by replacing \( F_i/\lambda + F_{cm}/\lambda \leftrightarrow F_i + F_e + F_c \) and \( \mathcal{H} \leftrightarrow T \). Generally, \( \mathcal{H} \) increases with height within the boundary layer and then decreases with height. This is because convective process brings the water vapor to the middle troposphere. The upward moisture flow inside the boundary is counter gradient to \( \mathcal{H} \), which produces the negative internal entropy production by enhancing the gradient of \( \mathcal{H} \). However, the negative value of \( \sigma_m \) is very small compared to other processes. The total internal entropy production is positive everywhere as shown in Fig. 3.

Above the boundary layer, the upward moisture flow is in downgradient to \( \mathcal{H} \), which produces a positive entropy production in the cyclone-located regions.

The term \( \sim (C - E) \ln \mathcal{H} \) shows the internal entropy production change due to the irreversible processes of phase change. The detailed physics of this term has been discussed in Li and Chylek (2012). It is found this term is very small for extratropical cyclone. The change in total internal entropy production is not obviously noticeable by including this term.

The total internal entropy production provides a comprehensive measurement of a cyclone, which is associated with all physical processes related to turbulent diffusion, convection, wind speed, wind stress, water vapor diffusion, phase change, temperature gradient, moisture gradient, etc.

As a cyclone moves from the eastern Pacific to the western coast of North America, strong internal entropy production is always occurring to such cyclones as shown above. Why does the cyclone not dissipate before it reaches North America? This is because of the presence of baroclinic forcing. As mentioned above, the cyclonic circulation due to baroclinicity can cause the conversion of potential energy to kinetic energy, which is necessary for the dynamic (wind) structure of a cyclone. Also, the poleward warm and moist front can cause latent heat release. Over the lifetime of a cyclone, the baroclinic force and dissipation balance to some extent. When a cyclone makes landfall, the poleward moist front and cyclonic wind structure are much reduced owing to the less-moist condition and larger land surface friction, and then the cyclone dissipates.
4. Advection of entropy flow and surface entropy supply

Generally, in a thermodynamic system, the internal entropy production dissipates the orderly structure and the external entropy production provides the boundary entropy supply. For a specific weather event, like a cyclone, the internal entropy production and the entropy boundary supply are not balanced at any moment, which leads to a nonzero entropy tendency. Applying the vertical integral to the entropy balance relation in (7), we have

\[
\int_0^{z_t} \frac{\partial (\rho s)}{\partial t} dz = -\int_0^{z_t} \frac{\partial \rho u_s}{\partial x_i} dz + \frac{E_h + F_L}{T} \bigg|_0 - \frac{\rho u_a T_3 a}{T} \bigg|_0 - R_u \frac{F_L}{T} \ln \theta \bigg|_0 + \int_0^{z_t} \sigma r dz + \int_0^{z_t} c_{pd} \theta \frac{Q}{T} dz, \tag{13}
\]

where \( z = z_t \) is the top of atmosphere. On the right-hand side of (13), the first term is the advection; the second and third terms are the boundary entropy fluxes due to sensible heat, latent heat, and surface momentum stress; the fourth term is the external entropy production due to surface evaporation under nonsaturation condition; and the fifth term is the internal entropy production. The fluxes related to convection are close to zero at the surface.

The internal entropy production as discussed in section 3, however, relies only on the resolved variables. Therefore, the entropy approach of cyclone study can be undertaken using the resolved physical quantities as well.

The top row of Fig. 9 shows the vertically integrated entropy advection at 0000 UTC 7 January and 1200 UTC 9 January. For an individual cyclone, it is interesting to note that the entropy advection is generally positive in the leading part and becomes negative in the trailing part. For a volume of \( V \) with surface \( \Omega \), we have

\[
-\int \int \int \frac{\partial}{\partial x_i} (\rho u_s) dV = -\int_\Omega \rho u_s d\Lambda_i, \tag{14}
\]

where \( d\Lambda_i \) is the surface integral element in the \( i \) direction. Thus, if the advection is positive, \( -\int_\Omega \rho u_s d\Lambda_i > 0 \Rightarrow \int_\Omega \rho u_s d\Lambda_i < 0 \), which means that the outgoing entropy flow is less than the incoming entropy flow. The net entropy flow is convergent to \( \Omega \). Similarly, the negative advection of entropy means that the net entropy flow is divergent to \( \Omega \). Therefore, entropy flow is convergent in the leading part of a cyclone and divergent in the trailing part. The detailed analysis shows that such convergence (divergence) is accomplished mostly through the horizontal directions. As a cyclone moves eastward, it takes up entropy flow in its leading edge and releases it in its trailing end. The net incoming and outgoing entropy flow is, however, not zero as the entropy advection could have contribution to the entropy budget as shown in (13).

In the bottom row of Fig. 9, the distributions of the surface entropy supply of

\[
\Sigma = \frac{E_h + F_L}{T} \bigg|_0 - \frac{\rho u_a T_3 a}{T} \bigg|_0 - R_u \frac{F_L}{T} \ln \theta \bigg|_0 + \int_0^{z_t} \sigma r dz + \int_0^{z_t} c_{pd} \theta \frac{Q}{T} dz, \tag{15}
\]

are displayed. In comparison with Fig. 2, the boundary supply of the entropy flux is highly correlated with the internal entropy production in both the Pacific and the Atlantic. The boundary supply of entropy flux is generally very weak in the eastern (leading) part of a cyclone and can even be negative. As shown in Fig. 3 in the warm sector east of the cyclone, near-surface air could be slightly warmer than SST, while in the cold sector west of the cyclone, near-surface air is much colder than the SST. Turbulence transfer is much enhanced in the cold sector owing to instability. It is known that the sensible or latent heat flow from ocean to atmosphere peaks over the cold air behind the cold front (e.g., Neiman and Shapiro 1993; Neiman et al. 1993). The surface entropy flux as shown in (13) provides an overall measurement as it brings together all the surface physical factors of sensible heat, latent heat, and surface evaporation, as well as wind stress and surface temperature.

The negative supply of entropy from the surface can partly compensate for the positive internal entropy production. In most storm-track regions, however, the surface entropy supply is positive. In a region, if both the internal entropy production and the external entropy flux are positive; like in the trailing part of a cyclone, the tendency of entropy in these regions will increase. From entropy balance as shown in (13), there exists a sink to limit the increasing entropy tendency. This is primarily achieved by entropy advection with strong divergence occurring in the trailing part, as shown in the top row of Fig. 9. Finally, we would like to point out that the nonzero entropy tendency along the storm track can also provide a useful diagnostic for cyclone identification.
5. A regional case study and cyclogenesis

In this section, cyclones are investigated in the area 20°–50°N, 110°–170°E. Along the Asian continent, the region warmed by the Kuroshio is the cyclogenetic region. Results for 3 days, 4–6 January 1995, are displayed in Fig. 10. Pacific cyclone numbers 2, 3, and 4 discussed in the previous sections underwent cyclogenesis during this period in this region.

In Fig. 10, the top two rows are the results at 1200 UTC 4 January. The first row gives the pressure anomaly, PV, and the temperature shear. The second row shows the surface entropy flux, the vertically integrated thermal dissipation, and dynamic dissipation. The third and fourth rows are as in the top two rows, but for 1200 UTC 5 January. The fifth and sixth rows are the results for 1200 UTC 6 January.

At 1200 UTC 4 January, there is a center of pressure anomaly in northeastern Japan, which is the number-2 cyclone. At this stage the number-2 cyclone is mature, as is clearly indicated by the surface entropy flux and thermal and dynamic dissipation. The contours of temperature shear are also close to the center of the pressure anomaly. The contour center of PV is located north of the pressure anomaly center. At the same time, there is a birth of a new cyclone near the East China Sea and the

![Fig. 9](image-url)
FIG. 10. (first row) The distribution of (left) the pressure anomaly with contour interval of 2 hPa, (middle) the vertically integrated PV with contour interval of $5 \times 10^{-8} \text{kg}^{-1} \text{m}^2 \text{s}^{-1} \text{K}$, and (right) the temperature difference with contour interval of 1.5 K. (second row) The distribution of (left) the surface entropy flux with contour interval of 0.25 W m$^{-2}$ K$^{-1}$, (middle) the thermal dissipation with contour interval of 0.025 W m$^{-2}$ K$^{-1}$, and (right) the dynamic dissipation with contour interval of 0.01 W m$^{-2}$ K$^{-1}$. The moment is 1200 UTC 4 Jan. The domain of each panel is 20°–50°N, 110°–170°E. (third row),(fourth row) As in first and second rows, respectively, but for 1200 UTC 5 Jan. (fifth row),(sixth row) As in first and second rows, respectively, but for 1200 UTC 6 Jan. Dashed lines are negative values.
Sea of Japan, which is a cyclogenesis region. This is the number-3 cyclone. At this incipient stage the signal of pressure anomaly is very weak.

In the traditional explanation for cyclogenesis, the divergence aloft in the mid- and high troposphere usually plays a key role. The details can be found in Wallace and Hobb (2006) and Holton (2004). At 1200 UTC 4 January, it is clearly shown that PV tilts toward the northeast near the East China Sea, indicating a PV advection from trough to ridge. This causes upper-level divergence and a starting of cyclogenesis. From nonequilibrium thermodynamics, what additional information for cyclogenesis can be obtained? First, at 1200 UTC 4 January the surface entropy flux is strong at the cyclogenesis region, which means the surface latent heat flux and sensible heat flux must be strong there. Another signal, having not been paid much attention to, is the temperature shear. There are two reasons for a large temperature shear at this stage: one is the higher surface temperature near the East China Sea and the Sea of Japan due to the Kuroshio, and another is the lower temperature in the mid- and high troposphere due to the divergence aloft. However, the corresponding thermal dissipation is not strong at this moment. Therefore, it is the existence of temperature gradient first, which induces the vertical turbulent diffusion and convection to produce a dissipation process later. In the early stage, the rotational structure of a cyclone is not fully established. There is no low-pressure anomaly shown in the location of a cyclone. Because of the weak dynamic structure, the signal of dynamic dissipation is also not apparent.

At 1200 UTC 5 January, the number-2 cyclone has developed more and moved eastward. The centers of low pressure, temperature shear, surface entropy supply, and dynamic dissipation are highly collocated. The number-3 cyclone appears to be more mature as the thermal dissipation and temperature shear become stronger. At this moment, there is a pressure anomaly center over northern Japan, as the cyclic rotation structure of the cyclone is gradually established. Also the dynamic dissipation starts to appear.

At 1200 UTC 6 January, the number-3 cyclone has moved off Japan. The surface entropy flux, thermal and dynamic dissipations, and temperature shear become much stronger and more collocated. The contour of PV corresponding to the number-3 cyclone appears clearly. At this moment, in the East China Sea and the Sea of Japan, there appear new centers of surface entropy flux and temperature shear. A new cyclone (number 4) starts to be born, although there is no pressure anomaly in this region yet.

In summary, based on the reanalysis data the nonequilibrium thermodynamics can provide some additional information for cyclogenesis. In a cyclogenesis region, the signals of the surface entropy flux and temperature shear generally appear first. With the development of a cyclone, thermal dissipation gradually becomes strong. The dynamic dissipation appears only when the cyclone has become mature. Although the disturbance in PV at higher altitudes is believed to be an important physical cause of cyclogenesis, the pattern change of PV in the East China Sea and the Sea of Japan is small. At least from the case study shown above, surface entropy flux and temperature shear are the most reliable early signals of cyclone.

6. Conclusions

In this paper, the physical characteristics of cyclones are explored with the aid of nonequilibrium thermodynamics based on the analysis of atmospheric entropy production. The entropy balance equation contains two parts: the internal entropy production, describing the dissipation process inside a system, and the external entropy production, describing the boundary entropy supply to the system. This study shows that dissipation is always present in a cyclone. There are three types of dissipation: thermal, dynamic, and moisture. At the incipient stage, the dissipation is mainly due to thermal contributions. In the mature stage, the dynamic dissipation starts to become important. The moisture is relatively weak. The temperature shear is an indication of thermal dissipation. It is found that centers of temperature shear and internal entropy production are generally well collocated.

In addition to the internal entropy production, the advection of entropy and the external entropy flux supply are investigated in terms of cyclone development. It is interesting that the entropy advection is positive (negative) in the leading (trailing) part of a cyclone. Analysis shows that the convergent/divergent parts are consistent with the warm/cold fronts of cyclones.

The regional study of cyclones in the western Pacific demonstrates the relationship among the low pressure, PV, temperature shear, surface entropy supply, and thermal and dynamic dissipations. At the initial stage of a cyclone, the low-pressure anomaly does not provide the clear signal of incipient cyclones. However, the surface entropy flux, temperature shear, and thermal dissipation provide the reliable signals, which can indicate cyclones in the cyclogenesis stage.

The methodology of atmospheric entropy provides several new ways to describe the evolution of a cyclone. It is shown that in most cases the physical feature of a cyclone is displayed more clearly by using these new approaches in comparison with the methods of pressure
anomaly, PV, and diabatic heating. The internal entropy production, external entropy supply, and entropy advection, etc. clearly reveal the different physical processes occurring at different stages of cyclone development.

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REFERENCES


